

173

1051-H-11

V3 N3 25/10/81

COMPLETED 144

NUREG/CR-4903
Vol. 2

Selection of Earthquake Resistant Design Criteria for Nuclear Power Plants - Methodology and Technical Cases

Methods for Introduction of Geological Data Into
Characterization of Active Faults and Seismicity and
Upgrading of the Uniform Risk Spectrum Technique

Prepared by J. G. Anderson, V. W. Lee, M. D. Trifunac

Structural and Earthquake Engineering Consultants, Inc.

Prepared for
U.S. Nuclear Regulatory
Commission

Handwritten signature/initials

199

BEST COPY AVAILABLE

NOTICE

This report was prepared as an account of work sponsored by an agency of the United States Government. Neither the United States Government nor any agency thereof, or any of their employees, makes any warranty, expressed or implied, or assumes any legal liability of responsibility for any third party's use, or the results of such use, of any information, apparatus, product or process disclosed in this report, or represents that its use by such third party would not infringe privately owned rights.

NOTICE

Availability of Reference Materials Cited in NRC Publications

Most documents cited in NRC publications will be available from one of the following sources

1. The NRC Public Document Room, 1717 H Street, N.W.
Washington, DC 20555
2. The Superintendent of Documents, U.S. Government Printing Office, Post Office Box 37082,
Washington, DC 20013-7082
3. The National Technical Information Service, Springfield, VA 22161

Although the listing that follows represents the majority of documents cited in NRC publications it is not intended to be exhaustive.

Referenced documents available for inspection and copying for a fee from the NRC Public Document Room include NRC correspondence and internal NRC memoranda; NRC Office of Inspection and Enforcement bulletins, circulars, information notices, inspection and investigation notices; Licensee Event Reports; vendor reports and correspondence; Commission papers; and applicant and licensee documents and correspondence.

The following documents in the NUREG series are available for purchase from the GPO Sales Program: formal NRC staff and contractor reports, NRC-sponsored conference proceedings, and NRC booklets and brochures. Also available are Regulatory Guides, NRC regulations in the *Code of Federal Regulations*, and *Nuclear Regulatory Commission Issuances*.

Documents available from the National Technical Information Service include NUREG series reports and technical reports prepared by other federal agencies and reports prepared by the Atomic Energy Commission, forerunner agency to the Nuclear Regulatory Commission.

Documents available from public and special technical libraries include all open literature items, such as books, journal and periodical articles, and transactions. *Federal Register* notices, federal and state legislation, and congressional reports can usually be obtained from these libraries.

Documents such as theses, dissertations, foreign reports and translations, and non-NRC conference proceedings are available for purchase from the organization sponsoring the publication cited.

Single copies of NRC draft reports are available free, to the extent of supply, upon written request to the Division of Information Support Services, Distribution Section, U.S. Nuclear Regulatory Commission, Washington, DC 20555.

Copies of industry codes and standards used in a substantive manner in the NRC regulatory process are maintained at the NRC Library, 7920 Norfolk Avenue, Bethesda, Maryland, and are available there for reference use by the public. Codes and standards are usually copyrighted and may be purchased from the originating organization or, if they are American National Standards, from the American National Standards Institute, 1430 Broadway, New York, NY 10018.

Selection of Earthquake Resistant Design Criteria for Nuclear Power Plants - Methodology and Technical Cases

**Methods for Introduction of Geological Data Into
Characterization of Active Faults and Seismicity and
Upgrading of the Uniform Risk Spectrum Technique**

**Manuscript Completed: September 1986
Date Published: May 1987**

**Prepared by
J. G. Anderson, V. W. Lee, M. D. Trifunac**

**Structural and Earthquake Engineering Consultants, Inc.
Arcadia, CA 91006**

**Prepared for
Division of Engineering
Office of Nuclear Regulatory Research
U.S. Nuclear Regulatory Commission
Washington, DC 20555
NRC FIN B8298
Under Contract No. NRC-04-83-022**

BLANK PAGE

ABSTRACT

This report reviews the physical and experimental bases for a quantitative relationship between earthquake occurrence rates and geological deformation rates. These relationships are well founded on mathematical statements of the elastic rebound theory, and well supported by observations.

The concept of uniform risk spectra of Anderson and Trifunac (1977) is generalized to include (1) more refined description of earthquake source zones, (2) the uncertainties in estimating seismicity parameters a and b in $\log_{10} N = a - bM$, (3) the uncertainties in estimation of maximum earthquake size in each source zone, and (4) the most recent results on empirical scaling of strong motion amplitudes at a site.

Examples of using the new NEQRISK program are presented and compared with the corresponding case studies of Anderson and Trifunac (1977). The organization of the computer program NEQRISK is also briefly described.

BLANK PAGE

TABLE OF CONTENTS

INTRODUCTION.....	1
APPENDIX I: A Review of Quantitative Procedures to Obtain Earthquake Occurrence Rates from Rates of Geological Processes: Application to the Eastern United States.....	I.1
APPENDIX II: Uniform Risk Spectra of Strong Earthquake Ground Motion: NEQRISK.....	II.1

BLANK PAGE

FOREWORD

This report presents partial results of a four-task study on the development of methodology and technical bases for the selection of earthquake resistant design criteria for nuclear power plants. This research program was sponsored by the U. S. Nuclear Regulatory Commission (USNRC) under Contract NRC-04-83-022.

Volume 1 of this report presents the results of Task III which include:

- IA) Attenuation of Modified Mercalli Intensity for small epicentral distance in California,
- IB) Frequency dependent attenuation of strong earthquake ground motion,
- IC) Preliminary empirical model for scaling Fourier amplitude spectra of strong ground acceleration in terms of earthquake magnitude, source to station distance, site intensity and recording site conditions,
- ID) Preliminary empirical models for scaling pseudo relative velocity spectra of strong earthquake acceleration in terms of magnitude, distance, site intensity and recording site conditions.

Volume 2 of this report presents the results of Tasks I and IV which include:

- IIA) A review of quantitative procedures to obtain earthquake occurrence from rates of geological processes: application to the eastern United States,
- IIB) Uniform risk spectra of strong earthquake ground motion: NEQRISK.

Finally, Volume 3 of this report presents the results of Task II which correspond to a review of a number of theoretical fault models to establish some of the expected characteristics of earthquake ground motion in the near-source region.

Overall technical guidance and review was provided by A. Murphy, technical representative of the USNRC for this research project.

INTRODUCTION

The current guidelines, on how the active faults are to be included in the development of the maximum vibratory ground motion for use in earthquake resistant design at a site (U.S.A.E.C., 1971, Reactor Siting Criteria: Appendix A: seismic and geologic siting criteria for nuclear power plant sites: Fed. Register V.36, 22601-22626), essentially require one to assume that the largest possible earthquake will occur during the expected life of the plant. For example, if a motion can be detected on a fault during the past 35,000 years, it is assumed that the largest earthquake supported by the fault system will occur during the next 50 years. Clearly this may exaggerate the actual probability that such an event will indeed occur. As a result, when an active fault is discovered during the site selection work or during a review of an existing plant site, the present approach (Appendix A) may lead to a too large emphasis on the risk assumed to result from an earthquake on the fault in question. It would, of course, be ideal if it were possible to predict the future earthquake shaking with deterministic certainty. However, keeping in mind numerous unknowns in the entire design process, it seems clear that such assumptions can sometimes lead to "unbalanced" risk predictions and thus may overshadow other perhaps more important considerations.

Starting with geological data, the seismicity for applications in seismic risk studies can be estimated. The rate at which seismic moment is released can be determined on a fault when the slip rate is known. It can also be estimated in a region of crustal convergence (without subduction) or divergence when the rate at which opposite sides of the zone are converging or the regional strain rate is known. Then, provided

all of the deformation is released seismically, by assuming the relative frequency of different sizes of earthquakes, the absolute frequency of events can be obtained.

This procedure has been used to estimate seismicity in Southern California. The estimates of the seismicity resulting from these slip rates are consistent with historical records of earthquake occurrences for Southern California taken as a whole. For smaller regions or individual faults in Southern California, the seismicity estimated from slip rates may differ from historical rates of seismicity by a factor of two or more. In the Western Basin and Range region, the historical seismicity is also consistent with an estimate for the strain rate. Because of this agreement in larger regions where many faults are involved, it is inferred that the geological data is also useful for studies of smaller regions, even though on this scale the model is difficult to test because of the too short historical record of earthquake occurrences.

The outcome of a seismic risk analysis depends critically on the description of the seismicity used as input (NUREG - 0406, Appendix E). "Seismicity" is used here to mean a complete description of the locations, sizes, and occurrence rates of earthquakes in a region. In a large enough region, the occurrence rate of earthquakes often seems well determined. But the seismic risk at a specific site is most sensitive to the distribution of earthquakes near that site, for example, within 25 to 50 km. On this scale, the historical record is usually too short to estimate the seismicity. Geological studies on this scale are routine, however, and their usefulness for studies of the seismic hazard is obvious. In this report we outline a framework for using these geological studies to obtain input to a seismic risk model.

Seismicity estimated from geological structure may be referred to here as the "geological seismicity." These estimates are for occurrence rates of earthquakes over a time interval of 10^3 to 10^6 years. This may be contrasted with the "historical seismicity," based on felt reports, which has a time scale of 10^2 to 10^3 years, and with "instrumental seismicity," which is known for a time scale of 10^1 to 10^2 years.

For the input to a risk analysis, a long term description of seismicity based on actual data is most desirable. This could be derived from an instrumental or historical record which is much longer than the average recurrence interval of large earthquakes on all the faults near the site. But since these recurrence intervals may often be 10^2 to 10^5 years, a detailed description of the long term seismicity is practically impossible to obtain from the historical or the instrumental seismicity. Significant periodicities of occurrence or an earthquake prediction could, at times, demonstrate that the long term estimate is too high or too low for risk calculations for a coming short time interval.

The geological record can often be interpreted to obtain fault slip rates or regional strain rates; these rates determine the geological seismicity. With this, several consistency checks are then possible: the known slip rates on major faults can be compared with slip rates derived from the rigid plate assumption in plate tectonic theory. These slip rates can be used in three ways to estimate the geological seismicity, and the geological seismicity can be compared with the historical seismicity. Agreement between the geological seismicity and the historical seismicity on the scale of Southern California implies that both methods give a good estimate for long term seismicity there (NUREG/CR-0689). Geological seismicity can also be estimated for the Basin and Range region on the basis

of suggested strain rates; there, too, it is consistent with the historical seismicity. It is thus suggested that the geological seismicity may be used to estimate the long term seismicity in smaller regions when the historical record is too short.

In this report, in Appendix I, we extend and test such techniques (NUREG/CR - 0689) to other selected test areas of the country and formulate this approach into a methodology which can be used as a tool in the selection of the earthquake design criteria at a site.

The use of the Uniform Risk Spectrum (URS) technique during the past several years has proved to be invaluable for independent testing and verification of the seismic design spectra developed through deterministic or other techniques. The balanced way in which the URS brings about the effects of the nearby small and sometimes frequent earthquakes, and large less likely and often distant earthquakes has been most helpful and quite stable. In spite of perhaps a common belief that the probabilistic calculations may be difficult to test, and that the results may seriously fluctuate as a function of the seismicity input and the assumptions in the model, we found the URS method to be very reliable and quite stable. For this reason, in this report (in Appendix II) we present a new and upgraded version of this technique to reflect the new developments and the more refined scaling and seismicity estimates.

At present the main features of the URS approach (NUREG - 0406, Appendix E) appear to be well accepted and in fact quite popular in advanced research and technical applications. Thus the main thrust of the current effort has been to focus on the development of simpler input/output characteristics of the program EQRISK (NUREG/CR - 0689, Appendix G) so that

more general and more detailed assumptions about the surrounding seismicity can be input into the URS calculations (see Appendix II).

REFERENCES

NUREG-0406 Vol. 1 and 2 (1977). Methods for Prediction of Strong Earthquake Ground Motion, U.S.N.R.C., Washington D.C.

NUREG/CR-0689 (1979). Methods for Prediction of Strong Earthquake Ground Motion, U.S.N.R.C., Washington D.C.

BEST COPY AVAILABLE

APPENDIX I

A REVIEW OF QUANTITATIVE PROCEDURES
TO OBTAIN EARTHQUAKE OCCURRENCE RATES FROM
RATES OF GEOLOGICAL PROCESSES:
APPLICATION TO THE EASTERN UNITED STATES

by

John G. Anderson

ABSTRACT

This paper reviews the physical and experimental bases for a quantitative relationship between earthquake occurrence rates and geological deformation rates. In brief, these relationships are well founded on mathematical statement of the elastic rebound theory, and well supported by observations. A quantitative test for the adequacy of a historical catalog, motivated from results of Smith (1976), is presented. The role of aseismic creep, the width of the seismogenic zone, the shape of the earthquake magnitude - occurrence rate curve (including maximum earthquake), and the limitations of geological observations to define deformation rates are the most important sources of uncertainty in the process.

Anderson (1986) presented evidence consistent with the hypothesis that erosion and deposition rates might control the seismic strain rate, or might determine the lower bound for the seismic strain rate. This is one of the few hypotheses that allow for prediction of earthquake rates in the eastern United States. This hypothesis might imply that earthquakes are without significance for tectonic deformation of the region and the widespread lack of surface faulting. It also might give a physical basis for estimates of the maximum earthquake, and allow for a (lower bound?) estimate of seismic hazard which is independent of the historical record of earthquakes.

TABLE OF CONTENTS

ABSTRACT.....	I.i
INTRODUCTION.....	I.1
REVIEW OF DEFORMATION RATE TECHNIQUES.....	I.3
RECURRENCE CURVES.....	I.27
THE PARAMETER $M_{\max}$	I.32
CONDITIONAL PROBABILITIES.....	I.36
SEISMIC STRAIN RATES IN THE EASTERN UNITED STATES.....	I.43
ACKNOWLEDGEMENTS.....	I.46
REFERENCES.....	I.47

INTRODUCTION

Along the active margins of the earth's tectonic plates, seismology and tectonics have developed good models for the geological processes that cause the earthquakes. These plates move relative to each other, separating at some margins, sliding past in other places, and on some margins one plate is consumed as it slides beneath another. Seismology played a crucial role in developing the plate tectonic model by recognizing the directions of slip between the plates and showing that these slip directions are those expected from the magnetic lineations on the sea floor. Later the relationship was made quantitative with demonstrations that the rates of earthquake occurrences on the plate margins match the rates that are expected from the time scale of the magnetic lineations.

These quantitative procedures can also be extended to aid the process of seismic risk analysis on a smaller scale. Especially in California and in Japan, the slip rate of quaternary faults is known far better than the earthquake activity rates on these faults. Consequently, the slip rates have been used to estimate fault activity rates.

Reviewing these methods, one comes to the question of whether they can be extended to the "stable" intraplate regions, such as eastern North America. With the experience of plate tectonics, it is easy to assume that the naturally occurring earthquakes everywhere contribute directly to plate motions, mountain building, volcanic activity, or other types of active deformation. Some earthquakes, of course, are triggered by man's activities, in which case it is easy to interpret the events as having minimal tectonic significance, although they do

in these cases convey information about the state of stress in the crust. For some, perhaps most of the intraplate earthquakes, also, it is necessary to consider the hypothesis that there is minimal significance for associated deformation of the earth. This might imply an inherent lack of localized, measurable faulting; also the condition that an earthquake source region lacks tectonic significance might place constraints on the earthquakes that can occur there.

This paper reviews the mathematical and experimental bases for a quantitative relationship between earthquake occurrence rates and geological deformation rates. It also reviews earthquake occurrence rate relations and the concept of an extreme earthquake magnitude that are essential for applying these techniques to seismic hazard estimates. After this is considered now one might apply the techniques to regions where earthquakes are the side effects of more passive geological processes.

REVIEW OF DEFORMATION RATE TECHNIQUES

Shallow earthquakes cause permanent deformation of the crust (Reid 1910). Intuitively, therefore, one expects that the occurrence rates of earthquakes are related to the rate at which the crust of the earth is deforming. This relationship can be made quantitative.

The permanent deformation caused by an earthquake determines a physical parameter called the seismic moment. Conveniently, the seismic moment of each earthquake is closely related to the magnitude. (The moment can also be estimated directly from seismograms.) Thus it is straightforward to use the instrumental catalogues of earthquakes for a region to estimate the average rate of deformation caused by earthquakes over the time interval covered by the catalog. In many active areas, where the earthquake occurrence rates are statistically meaningful, they are consistent with deformation rates obtained from independent geological or geophysical investigation. Conversely, if one knows the geological rate of deformation, it is not difficult to determine occurrence rates of earthquakes which will accomplish the deformation.

The method outlined above has been applied most extensively in a situation where the deformation consists of simple shear on a fault, and the slip rate is the measure of the deformation rate. The second environment in which these relationships have been exploited is regional deformation in which individual faults are ignored and the overall strain rate is obtained.

Faulting: Moment Rates and Slip Rates

It is useful to describe the size of earthquakes in terms of the seismic moment instead of magnitude. The seismic moment of earthquake k is:

$$M_0 = \mu \sum_k u_k \quad (2.1)$$

where μ is the shear modulus, $\sum_k$ is the area of a fault which ruptured during an earthquake, and u_k is the spatial average slip on the fault (over the area $\sum_k$) during the earthquake. The moment is the fundamental parameter which describes static aspects of earthquakes. Aki (1966, 1967) showed that the long period spectral density of body and surface wave displacement is proportional to M_0 .

Aki (1966) also employed an M_0 estimate to obtain an estimate of the average slip during the Niigata earthquake of June 16, 1964 ($M=7.5$). The shear modulus was estimated from the shear velocity (β) and density (ρ) of the crust:

$$\mu = \rho \beta^2 \quad (2.2)$$

The area was obtained as a product of the length of the aftershock zone and the width of the fault, estimated from the epicentral depth.

The assumption about homogeneity of the crust (μ independent of depth) persists to even the most recent studies. It would be interesting to evaluate the effect of this assumption.

During the late 1960's, the relationship between seismic moment measured from geological observations and as measured from the seismic radiations was verified (Brune and Allen, 1967; Brune 1968; Aki 1966). Subsequent studies have consistently verified this observation.

The statistical relationship between M_0 and various magnitude scales was also established in the 1960's and refined subsequently. Typically this relationship is expressed in the form:

$$M_0 = 10^{C+dM} \quad (2.3)$$

Brune et al. (1963) defined a parameter, AR, as the sum of the areas of long period surface waves on a certain kind of three-component seismograph, and Wyss and Brune (1968) employed this parameter to estimate the seismic moment of over 250 California earthquakes. This was added to the statistical correlation described above.

Brune and King (1967) studied the relative spectral amplitudes of 20 sec and 100 sec surface waves, as a function of magnitude, and Brune (1968) employed these, plus seismic moments measured from the geological observations of Brune and Allen (1967) to derive a moment - magnitude curve for the M_s scale.

Hanks and Boore (1984) have prepared a summary of many existing relations, and also discuss their relationship to seismic scaling laws. It is worthwhile to mention one other relationship, on which Hanks and Kanamori (1978) defined a moment magnitude by:

$$M = \frac{1}{d}(\log M_0 - c) \quad (2.4)$$

where $c = 16.0$ and $d = 1.5$. This relationship, which resembles several empirical correlations, has been used extensively.

This level of development in the studies of seismic moment allowed Brune (1968) to compare the rate of seismic slip with geological slip rates. To accomplish this, we let kM_0 be the moment of an earthquake

indexed by the integer k , and sum M_0 over all earthquakes which have occurred on a fault segment. The average slip from this suite of earthquakes is then:

$$U_f = (\sum_k M_0) / \mu \Sigma_f \quad (2.5)$$

where now Σ_f is the total area of the fault. If all these earthquakes occurred in the time interval of duration T , one obtains the average slip rate during that interval, S_T :

$$S_T = U_f / T \quad (2.6)$$

Brune (1968) and Davies and Brune (1971) demonstrated this technique. The paper by Brune demonstrated several useful concepts. The first is that the largest earthquakes dominate in the contribution to total slip. This is not obvious because small earthquakes greatly outnumber large ones. In a typical region a decrease of one magnitude unit leads to about a 30 fold decrease in the seismic moment (see equation (2.4)). Thus the large earthquakes dominate in causing permanent deformation.

A second important result observed by Brune (1968) is that the width of the seismogenic zone varies considerably from one location to another. On mid-ocean transform faults, the seismogenic layer may be 5 km wide, in contrast to continental transform faults, such as the San Andreas system, where it is 10 to 20 km wide. Brune suggested that this is related to different crustal structures and temperature profiles. Brune also suggested that a thinner seismogenic zone may be associated with a smaller maximum magnitude earthquake.

The third major point of the Brune (1968) paper concerns the role of aseismic slip, or creep. After calculating the seismic slip rate S_T , the total slip rate for the interval is the sum of S_T and the creep rate. Alternatively, the creep may be regarded as reducing the effective width of the seismogenic zone, thus reducing Σ_f and increasing U_f and S_T .

Finally, Brune (1968) emphasized the importance of the short sampling interval (50 to 100 years) contrasted with the time required to obtain S_T reliably (perhaps 1,000 years for a plate boundary fault). A rule of thumb is that one needs to sample several cycles of the maximum earthquake on a fault before S_T will resemble the slip rate determined from geological studies.

Davies and Brune (1971) further established the method of comparing slip rate S_T with the plate-tectonic slip rate. They compared these values for the major seismogenic zones worldwide, and demonstrated agreement. They recognized the possibility that there have been significant changes in the moment release rate on a worldwide scale in the historical time interval. This must be taken as a warning, since when the geographical extent of the study is decreased, the time interval required to obtain statistical stability goes up.

Reichle et al. (1976) and Reichle and Reid (1977) employed seismic moment rates in a study of the seismicity of the Gulf of California. The paper by Reichle et al. (1976) concluded that the width of the seismogenic zone is 3 to 5 kilometers in this region. This is so narrow that earthquakes above magnitude 5 or so probably rupture the full width of the seismogenic zone. This paper found that for this

particular area, a relatively short time interval (1918 to 1974) yields a sufficient rate of earthquakes to reconcile seismicity and geological slip rate with a fairly high degree of confidence.

Additional studies which have employed seismicity to estimate deformation rates include Greensfelder et al. (1980), Main and Burton (1981), Wesnousky et al. (1982).

The next step in the application of geological data to the seismicity problem comes with the recognition that an occurrence rate relationship can be constrained by the geological data. Suppose $N(M)$ represents the number of earthquakes per year with magnitude greater than M . There are many proposals in the literature for the shape of $N(M)$. Regardless of the shape, the following relationship must hold:

$$\dot{M}_0 = \int_{-\infty}^{\infty} M_0(M)n(M)dM \quad (2.7)$$

$\dot{M}_0$ is the cumulative moment which is released per year in the region with earthquake occurrence rate described by $N(M)$. The quantity $n(M)$ is an occurrence rate density function:

$$n(M) = - \frac{dN(M)}{dM} \quad (2.8)$$

The negative sign is introduced because $N(M)$ represents the integral of $n(M)$ from M to infinity.

The function $M_0(M)$ might be described by equation(2.3). Weichert and Hyndman (1983) and Hyndman and Weichert (1983) have discussed how the scatter about the mean in the function $M_0(M)$ affects results. Including the scatter will tend to cause $\dot{M}_0$ to be larger than when it is neglected in the integral in equation(2.7).

On a fault, the moment rate $\dot{M}_0$ is equal to:

$$\dot{M}_0 = \mu \Sigma_f S \quad (2.9)$$

Equation(2.7), with equation(2.9), is very similar to the process described by equations(2.5) and (2.6). The difference is that the magnitude distribution, as sampled by the historical record of earthquake occurrences, has been replaced by a distribution function which is intended to represent that data. But when S is known from independent geological data, equation(2.9) may also be used to determine $\dot{M}_0$.

A typical occurrence rate relationship contains three parameters. For example, the relation:

$$N_1(M) = 10^{a-bM} H(M_{\max} - M) \quad (2.10)$$

from Gutenberg and Richter (1954), has unknown parameters a , b and $M_{\max}$. If $\dot{M}_0$ is known, then by means of equation(2.7), expressions can be derived to express any one of these unknown parameters as a function of the other two. Anderson and Luco (1983) have carried out an examination of this mathematical procedure for three forms of $N(M)$. Main and Burton (1981) have evaluated equation(2.7) for two additional descriptions of earthquake occurrence rates.

Smith (1976) applied these constraints to estimate $M_{\max}$. The result, for $N_1(M)$ is, in the notation of this paper:

$$M_{\max} = \frac{\log[(\frac{d}{d-b}) T M_c] - c}{d} \quad (2.11)$$

The parameter $T M_c$ represents the cumulative slip for all earthquakes in the time interval T . Smith obtains this parameter from the

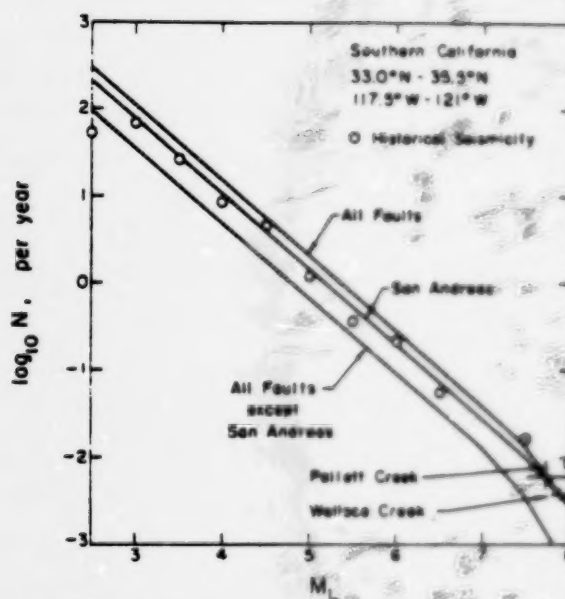


FIG. 3. Observed and predicted occurrence rates of earthquakes for the sub-region bounded by 33.0°N and 35.5°N, and 117.5°W and 121°W. Predicted rates are obtained by integrating equation (5) over intervals of $\frac{1}{2}$ magnitude width. The steeper slopes at larger magnitudes result when faults stop contributing to the total beyond their cut-off magnitude.

Figure 2.2. Observed and predicted earthquake occurrence rates for a portion of Southern California. Predicted rates are obtained from estimates of slip rates in many of the active faults in the region. (From Anderson, 1979).

boundary is around 6.5 cm/yr; if the assumptions of Wang et al. on shear modulus, length, and width of the seismic zone are valid, and if the historical record is long enough, then only about half of the convergence is occurring seismically. This conclusion that there is significant aseismic slip was also reached by Singh, et al. (1981).

The work of Davies and Brune (1971) has been repeated by Peterson and Seno (1984). Peterson and Seno worked with the moment release rate, defined as moment rate per kilometer of the subduction. With the research results since 1971, they were able to employ better estimates of the seismic moment of most earthquakes, and better plate tectonic models for convergence rates of the plate. They find that for most plates, the fraction of seismic slip is less than 50%. A particularly puzzling conclusion of the Peterson and Seno paper is that the moment release rate is not well correlated with the rate of convergence of the two plates because of overriding variations in the fraction of seismic slip. The moment release rate is better correlated to the absolute velocity of the subducted plate in the hot-spot frame of reference. Their analysis leads to the additional conclusion that the fraction of seismic slip decreases with increasing age of subducted material around the margins of a single plate, but for different plates the trends are not superimposed. These are intriguing trends with potentially great tectonic significance. Research is needed to better understand and predict the fraction of seismic slip.

Faulting: Comments on Selected Applications

Papastamatiou (1980) has considered two additional applications for the constraint imposed by slip rate. The first is to combine with

product $\mu \Sigma_f u_g$, in which u_g is the fault offset observed from geological observations.

The parameter b (b -value) was obtained from small earthquakes. The parameter a enters explicitly by the assumption that T equals (or exceeds) the recurrence time of the maximum earthquake.

An interesting suggestion in Smith's (1976) paper is for a procedure to estimate how long a time interval, T , is required to obtain a sufficient sample to establish earthquake occurrence statistics. The procedure, when quantified, is in the form of an uncertainty principle:

$$L_z T > K_s \quad (2.12)$$

where L_z is the length of the seismicity zone, T is the observation period, and K_s is a constant to be determined. From global seismicity, K_s would be of the order of 4×10^6 km - years (80,000 km x 50 years). An uncertainty principle with form (2.12) can be developed from equation (15) in Anderson and Luco (1983). We obtain

$$L_z T(M) = G' \frac{M_c(M_{\max})}{\mu W_z S} \quad (2.13)$$

where G' is a dimensionless parameter which depends on the shape of the recurrence curves and on $M_{\max} - M$. For earthquake distributions with small b -values or a deficit of small earthquakes relative to the number of large earthquakes (measured by equation (2.10)), and for $M_{\max} - M \sim 0.5$, we find $1 < G' < 2$. Therefore one can write

$$K_s \approx 10 M_0(M_{\max}) / \mu W_z S$$

where the coefficient 10 provides for 5 to 10 repeats of near-maximum earthquakes.

Table 2.1 gives numerical estimates for K_S for some selected seismic zones. For global seismicity, if one takes $M_0(M_{\max})$ to be 10^{30} dyne-cm, W_z to be 100 km, μ to be 6×10^{11} dyne-cm, and S to be 4 cm/year, one obtains $K_S = 4 \times 10^6$ km-years in agreement with the empirical estimate inferred from Smith (1976). For regions with smaller slip rates, the smaller value of S will tend to increase K_S , but $M_{\max}$ also may be decreased significantly. Since both slip rates and $M_0(M_{\max})$ may vary by orders of magnitude, the overall effect is not straightforward to anticipate. The uncertainty in any estimate of K_S is probably at least half an order of magnitude. However, reasonable estimates for various regions (Table 2.1) indicate that for regions on the dimension of a typical seismic risk study (a linear dimension on the order of 10^2 km), the requirement on T_S generally will exceed the historical period of observation by orders of magnitude.

In Table 2.1, the smallest value of K_S is 3×10^4 km-years for the Gulf of California. Reichle et al. (1976) studied a 56 year record of instrumentally recorded seismicity of the region, and this record supported the geological estimate of slip rate in the Gulf within the uncertainties of the data, and at a fairly high level of confidence. It is, therefore, interesting to examine whether the inequality given in equation(2.12) is satisfied in the region. Equation(2.12) is either satisfied or nearly satisfied for the region taken in its entirety. Reichle et al. also divided the Gulf of California into three sub-regions, in which equation(2.12) is not satisfied and in which slip rate estimates determined from the seismic history are not as well defined.

TABLE 2.1
VALUES OF K_s FOR SEVERAL FAULTING ENVIRONMENTS

<u>Region</u>	μ (dyne/cm ²)	W_z (km)	S (mm/yr)	(M_{max})	$M_0(M_{max})$ (dyne-cm)	K_s (km-years)
Worldwide	6×10^{11}	100	40	9.6	2.5×10^{30}	1×10^7
San Andreas	3×10^{11}	20	40	8.0	1.0×10^{28}	4×10^5
San Diego	3×10^{11}	15	2	7.0*	3.2×10^{26}	4×10^5
Charleston	3×10^{11}	20	0.1*	7.0*	$1.6 \times 10^{27+*}$	3×10^7
Gulf of California	3×10^{11}	5	55	7.0	2.8×10^{26}	3×10^4
Middle America Trench	3.3×10^{11}	85	65	8.0	1.0×10^{28}	5×10^4

*Speculative - used for illustrative purposes only

+Based on Nuttli (1983) relationship between m_b , M_0 for mid-plate earthquakes

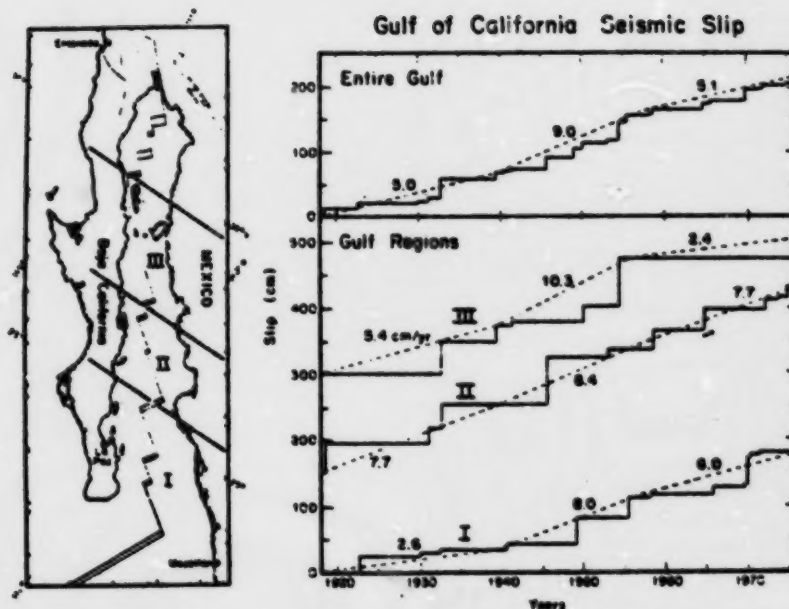


FIG. 11. Cumulative seismic moment (in 10^{20} dyne-cm). Map at left shows division of the Gulf into three regions. Lower three curves are cumulative seismic moment for each region (only $M_s \geq 5.5$ are plotted); Upper curve is for all three regions. Average slip rates (indicated by straight-line segments) are corrected by a factor of 1.6 to account for events down to $M_s = 3$.

Figure 2.1. Map of the Gulf of California, cumulative slip estimated from the instrumental catalog of seismicity, and interpretations of average slip rate in cm/year for the entire Gulf and for three sub-regions (from Reichle, 1976).

Figure 2.1, from Reichle et al. (1976) shows a map of the region and cumulative slip and determinations for three sub-regions, for the entire Gulf. The fault length for the entire Gulf is 880 km, implying 40 years of seismicity data is sufficient to define the seismicity. (Within the uncertainty of parameters, the time required to satisfy equation (2.12) may be 60 years.) For time intervals of 20 years, Figure 2.1 shows that Reichle et al. found room to interpret periods of differing slip rates, but an average over 40 to 60 years would leave much less room for ambiguity. The three sub-regions, which individually fail to satisfy the inequality by a factor of about three, also do not show a slip rate which is as well defined as it is for the region as a whole. For example, the average slip rate in region II is 2.4 times that in region I. Of course, it is possible that this discrepancy is the result of physical causes rather than the result of insufficient sampling, but it is equally plausible that the discrepancy will disappear as the seismic history of the region is compiled over the next century. Thus, the data for the Gulf of California are consistent with equation (2.12), with the seismicity constant defined by equation (2.13).

Molnar (1979) and Anderson (1979) independently published papers in which the constraints dictated by equation (2.7) were employed to estimate the occurrence rates of earthquakes. Molnar employed the occurrence rate relation equation (2.10) and emphasized worldwide applications at major plate boundaries and uncertainties in the applications. Anderson employed the occurrence relation:

$$n_2(M) = 10^{a-bM} H(M_{\max} - M) \quad (2.14)$$

and emphasized a more detailed application to within the Western United States.

The difference between equations (2.14) and (2.10) is subtle, as the expression for the density function in equation (2.12) is identical to the expression for the cumulative function in equation (2.10). When the cumulative function in equation (2.10) is differentiated, a Dirac delta function is introduced at $M_{\max}$ which is absent in equation (2.14). Weichert and Hyndman (1983) consider this delta function to be unphysical. The difference introduces an uncertainty of about 1.5 to 2 in the earthquake occurrence rates derived to be consistent with slip rate (Weichert and Hyndman, 1983; Hyndman and Weichert, 1983; Anderson and Luco, 1983).

One of the examples considered by Molnar was the estimation of occurrence rates of great earthquakes worldwide. Table 2.2 shows a study of this problem. In addition to the parameters selected by Molnar, we show the predictions for two other choices of model parameters. These and additional calculations show that within fairly broad limits on the shape of the occurrence relation, the b-value, and $M_{\max}$, it is possible to get reasonable agreement between the observed occurrence rates and a model. Agreement at the level cited by Molnar (1979) thus cannot be construed as support for that particular occurrence model, but rather only a demonstration that it is plausible. The physics of the problem is working through equation (2.7) to give even poor models for $N(M)$ their best possible exposure.

Molnar (1979) has carefully discussed the uncertainties which enter into these calculations. The parameters to be estimated include b , μ , L_f , W_f , S_g , $M_{\max}$ [or $M_o(M_{\max})$] and the parameters in the magnitude-moment

TABLE 2.2
APPLICATION EXAMPLE (from Molnar, 1979)

FAULT ZONE GEOMETRY

L_f	=	33,000 km	Occurrence relation:	N_1
W_f	=	100 km		
μ	=	7×10^{11} dyne-cm		
S_g	=	75 mm/yr		

Model 1: N_1 with $b = 1.0$ and $M_{\max} = 9.53$ (Molnar choices)

Model 2: N_2 with $b = 0.8$ and $M_{\max} = 10.0$

Model 3: N_1 with $b = 0.8$ and $M_{\max} = 9.6$

M_0	Corresponding M	Events/year $\geq M$			
		Data	Model 1	Model 2	Model 3
2×10^{30}	9.53	0.020	0.029	0.21	0.36
2×10^{29}	8.87	0.10	0.14	0.11	0.12
2×10^{28}	8.20	0.34	0.63	0.40	0.42

The following represents a test of several combinations of $M_{\max}$ and b for agreement with observations at the levels of M_0 given above. A response "YES," indicates agreement within a factor of two at all three levels, can be obtained by proper choice of the remaining free parameter, a .

$M_{\max} b$	N_1			$M_{\max} b$	N_2		
	1.0	0.8	0.2		1.0	0.8	0.2
9.53	Yes	No	No	9.53	No	No	No
9.60	Yes	No	No	9.60	?	No	Yes
9.80	Yes	Yes	No	10.0	Yes	Yes	No

relationship. In regions where the tectonics are reasonably well understood (usually associated with high or moderate seismicity), Molnar (1979) suggested that these parameters do not pose a serious problem. All are susceptible to measurement except $M_{\max}$. The tectonics in these situations are likely to allow an estimate for M_0 ($M_{\max}$) to within a factor of 5 from the fault geometry.

Greater uncertainties result from our ignorance of tectonic processes which are operating in some locations, from the shape of $N(m)$, and from the role of fault creep. All things considered, Molnar concluded the uncertainty in the frequency of occurrence of events with $M > 6$ to be at least a factor of 3 and probably 5. On the other hand, he suggested that careful geological studies have the potential to greatly reduce these uncertainties.

The paper by Anderson (1979) emphasized the application of the procedure to two regions: Southern California and the Basin and Range province. For Southern California, he employed geological data to try to estimate slip rates on all major faults, and employed fault length to estimate $M_{\max}$. From these, Anderson calculated occurrence rates for comparison with the observed rates. The result, shown in Figure 2.2, is satisfactory agreement.

The problem of aseismic creep is not trivial to deal with. For example, recently, McNally and Minster (1981) and Wang et al. (1982) have applied the technique of summing the seismic moments of earthquakes along the Middle America Trench adjacent to Mexico. Based on more complete data, Wang et al. inferred the seismic slip rate of 3.4 cm/yr. From plate tectonics, an average rate of convergence along this plate

an earthquake source model and estimate the probability of local ground rupture. Anderson and Luco (1983) have elaborated on the mathematical aspects of this possibility.

Papastamatiou's other new application involves estimating $M_{\max}$ from the current level of occurrence of small earthquakes. Anderson and Luco (1983) have also elaborated the mathematical aspects of this approach. This procedure carries with it the implicit assumption that instrumental seismicity at small magnitudes equals the long term average rate. This assumption may be valid for some faults, but in the general case there is no reason to assume its validity. It is plausible that as strains accumulate in a region toward the point where a major earthquake will occur, these higher regional strains will affect the occurrence rate of small to moderate sized earthquakes. For example, with typical parameter values, an error in occurrence rate of small events equal to a factor of 10 corresponds to an error in $M_{\max}$ of 2 magnitude units. Papastamatiou (1980) recognized these uncertainties but also considered that in some cases this procedure might be able to place some constraint on the probabilistic seismic risk.

As was mentioned previously, aseismic slip may decrease overall agreement between geological estimates and historical or instrumental seismicity. For a fault with total slip rate S_g and creep rate C , the slip rate entering the calculation of moment rate $\dot{M}_0$ should be S_s (Wallace, 1970):

$$S_s = S_g - C . \quad (2.15)$$

The uncertainty arises when S_g predicts an occurrence rate which is larger than the historical estimates. It is tempting to attribute such

a discrepancy to aseismic creep, but the alternative explanation is that in reality the sample interval is too short. In ten regions considered by Hyndman and Weichert, geological and seismological rates agree, but along the convergent plate margin of Oregon, Washington and southern British Columbia the tectonic models imply $S_g = 40$ mm/yr, compared with $S_s \approx 3$ mm/yr from seismicity for the interval 1900 to 1980 (Hyndman and Weichert 1983). Heaton and Kanamore (1984) have brought together evidence from plate tectonics that this particular discrepancy might be caused by accumulation of elastic strains. The multi-disciplinary lines of evidence which they present represent a reasonable approach for discriminating in other situations. The study by Anderson (1984) compared observed and predicted occurrence rates for 57 faults in California. The rates generally agree to within a factor of 10.

Volume Deformation: Theory

In this section, the case of volume deformation caused by earthquakes is considered. The earthquakes continue to occur on faults. However, now many faults, possibly with a wide range of fault dimensions and orientations, contribute to the deformation. Thus we are looking at deformation on a scale which is large compared to the fault spacing. The concepts of this section have been applied to central Asia, where master faults have been identified, and to the Basin and Range of the western United States, where numerous individual faults are recognized but none are seen to dominate the deformation. Much of the background seismicity in the eastern United States might be attributed to this sort of mechanism.

The spacing of the faults compared to changes in the attenuation function will determine what seismic risk techniques may be applied. Where faults are unknown or closely spaced, or where amplitudes of ground motion attenuate slowly with distance so that source to station distance is not critical, then a diffuse zone of seismicity consistent with the regional deformation rate can suffice to estimate the hazard of ground shaking.

Up to this point, the seismic moment has been treated as a scalar quantity. However, in any one earthquake the orientation of the fault and the slip direction are important determinants of the nature of permanent ground deformation. It has been implicitly assumed that these quantities are consistent from one earthquake to the next. In dealing with an ensemble of faults, that assumption is no longer valid. Therefore, one defines a 3 by 3 moment tensor to handle this more general situation.

Let $\underline{n}$ be a unit vector perpendicular to the fault plane of a specific earthquake and let $\underline{d}$ be a unit vector in the direction of fault slip for this earthquake. The scalar moment, M_0 , is defined as before. The symmetric moment tensor, M_{ij} , is defined by Kostrov (1974) by:

$$M_{ij} = M_0(d_i n_j + n_i d_j) \quad (2.16)$$

Molnar (1983) has defined an asymmetric moment tensor, M_{ij} , by:

$$M_{ij} = M_0 d_i n_j \quad (2.17)$$

These two definitions apply in different situations. When the region of interest is large compared to individual faults within the region, then

equation (2.16) is the correct one to apply. The average strain in the volume caused by the earthquake, which is denoted by the tensor $\underline{e}$ with components e_{ij} , is:

$$e_{ij} = M_{ij}/(2\mu V) \quad (2.18)$$

If the fault intersects the boundary of the region, so that the strains caused by end effects of the fault may be neglected, then it is more accurate to describe the deformation of the region as simple shear rather than pure shear, and the average strain of the volume becomes:

$$e_{ij} = M_{ij}/(\mu V) \quad (2.19)$$

Equation (2.19) provides a transition between the equations which are appropriate to slip on a fault and equations (2.16) and (2.18) for pure shear volume deformation. For most of the eastern United States, the simple shear situation is probably not important to consider. Therefore, equations (2.16) and (2.18) will be used in the following discussion.

When an ensemble of earthquakes is considered, the elements of the strain tensor caused by each are added to find the total average strain in the volume (Kostrov 1964). Subsequently, the strain tensor of the region can be rotated to find principal component directions and amplitudes. Chen and Molnar (1977) followed such a procedure for seismicity in Asia.

Anderson (1979) and Papastametiou (1980) suggested that it may also be useful to relate the total moment released by the earthquakes to the strain rate along the principal axis of tectonic compressions or deformation. The relation which is derived is, from Anderson (1979):

$$\dot{M}_0 = 2.67\mu V\dot{\epsilon} \quad (2.20)$$

Equation (2.20) has been expressed in terms of moment rate and strain rate ($\dot{\epsilon}$), rather than total moment and total strain. Alternative values for the coefficient, instead of 2.67, are a theoretical estimate of 2.5 (Papastametiou, 1980), 2.08 derived from data in southern Utah (Doser and Smith, 1982); and 2.41 derived from data in northern Utah (Doser and Smith, 1982). The coefficient 2.67 suggested by Anderson (1979) was based on the data of Chen and Molnar (1977) for central Asia. Anderson (1979) also obtained a value (3.14) from a theoretical procedure which would tend to overestimate the coefficient.

Given $\dot{M}_0$ from (2.18), (2.19), or (2.20), it can be used as a constraint on the occurrence rate relation of earthquakes in the same manner as it was used in section (2.1).

Volume Deformation: Examples

Chen and Molnar (1977) and Molnar and Deng (1984) have employed the seismicity of Asia to determine how parts of that continent are deforming. These papers, then, apply the principles of the previous section in a way similar to the contributions of Brune (1968) in estimating slip rates on faults from seismicity.

Anderson (1979) estimated earthquake occurrence rates in the Basin and Range province from geological estimates of strain rates. Figure 2.3 is taken from Anderson (1979), and shows that the rates are consistent.

Greensfelder et al. (1980a,b) carried out a more detailed investigation of the Basin and Range province. Moment rates were inferred

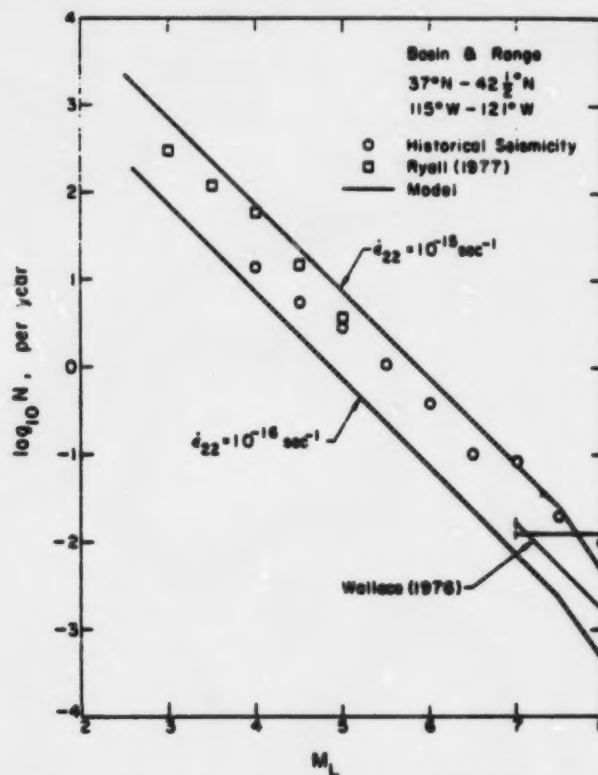


FIG. 4. Observed and model seismicity for the western basin and range region bounded by 37°N to 42.5°N and 115°W to 121°W . The historical seismicity is from Table 5. Data from Ryall (1977) are for years 1970 to 1974. Model uses maximum magnitude of 8.0 and b-value of 1.0, and is shown for extreme estimates of strain rate in the basin and range region from Lawrence (1976).

Figure 2.3. Comparison of observed earthquake occurrence rates and predictions from the range of geological strain rates which apply in the basin and range region of Western United States. (From Anderson, 1979).

from geological estimates of strain rates for seven sub-provinces, and compared with moment rates derived from earthquake occurrence rates. They found that the seismic moment rates were bounded by a range of 0.2 to 1.0 times the prediction from geology. Several explanations for the discrepancies are available, including aseismic strain release, uncertainty in geological estimate of strain rates, uncertainty in parameter estimates b , $M_{\max}$, etc., and a gradual slowing of strain rates with time. Also, the discrepancy is of the type which would result from an historical record of seismicity which is too short to observe large events.

Doser and Smith (1982, 1983) employed these techniques in their studies of seismicity in Utah and Wyoming. They calculated moment rate tensors from seismicity in the same manner as Chen and Molnar (1977). These agree with moment rates derived from geological observations.

Wesnousky et al. (1982) employed these techniques in their study of the seismicity of Japan. They calculated strain rates from the seismicity and from fault slip rates, and found that these are essentially in agreement.

Summary

This section has pointed to several of the studies which have tested the consistency of seismic moment rates and geological slip rates. We conclude that the physical basis for this consistency, and also the empirical evidence which supports this consistency, is strong. Therefore, it is reasonable to employ this consistency to obtain constraints on the earthquake occurrence rate wherever geological data will allow estimates of slip or strain rates.

RECURRENCE CURVES

To calculate earthquake occurrence rates from the moment rate, it is required to assume a shape for the recurrence curve, and then scale that shape according to the constraint provided by equation (2.7). Equations (2.10) and (2.14) offer two specific shapes of these recurrence curves. Many more shapes have been proposed based on a variety of theoretical and observational techniques. It is beyond the scope of this discussion to consider the history of this subject thoroughly.

Recurrence rates can be written as a cumulative distribution:

$$N(M) = \text{number of earthquakes with magnitude} \\ \text{greater than or equal to } M, \quad (3.1)$$

as a density function:

$$n(M) = -dN(M)/dM, \quad (3.2)$$

or as an incremental occurrence rate:

$$N^I(M) = N(M - \Delta M^I/2) - N(M + \Delta M^I/2) \quad (3.3)$$

which gives the number of events in a magnitude band of width ΔM^I centered at M . When $n(M)$ is reasonably smooth, and for sufficiently small ΔM^I , one can write:

$$N^I(M) = n(M)\Delta M^I. \quad (3.4)$$

Equations (2.10) and (2.14) both employ an exponential form, except that (2.10) attributes this exponential form to the cumulative distribution, while (2.14) attributes it to the density function. Anderson and Luco (1983), Weichert and Hyndman (1983), and Hyndman and Weichert

(1983) point out that this distinction can have important consequences in occurrence rate predictions.

In discussing the shape of the recurrence curve, it is useful to distinguish three cases.

Case I. Recurrence curves which apply to a region consisting of perhaps several faults. Examples are the world, Southern California, Japan, etc.

Case II. Recurrence curves which apply to a single fault. This fault may be of considerable length, so that several large earthquakes may occur in the process of the fault rupturing once along its length. Examples are the Alaska subduction zone, the San Andreas fault, etc.

Case III. Recurrence curves which apply to a single segment of a single fault. Examples are the Parkfield segment of the San Andreas fault, or the Hobbie Creek (Provo) segment of the Wasatch fault.

The discussion presented here will also employ the distinction of Scholz (1982) between small and large earthquakes. Following Scholz, the term large earthquake will be used here to mean specifically an earthquake which ruptures the full width of the seismogenic zone, for example from the surface to the depth where failure occurs by creep. A small earthquake is one which is entirely confined within the seismogenic zone. It is clear that these two cases have differing boundary conditions. Therefore, if there is a discontinuity in the earthquake recurrence relation, it would be expected at the transition from small

to large earthquakes. This transition will occur at different magnitudes for different faults.

Much of the early evidence for an exponential magnitude distribution is derived from regional seismicity (Case I). For small earthquakes, there are several theories which give the exponential distribution of earthquakes in magnitude. Examples include Caputo and Console (1980) and Andrews (1980). Wesnousky et al. (1983) demonstrated the plausibility of the hypothesis that for large earthquakes such curves result from a distribution of faults of different lengths, with each fault causing only one size of earthquake. For a single fault segment (Case III), they proposed that an exponential distribution is obeyed in foreshocks and aftershocks, but that this exponential distribution is truncated at the magnitude M^a of the largest aftershock. The maximum magnitude earthquake is represented by an isolated spike on the recurrence curve at a characteristic magnitude representing the size of the main shock. Some geological observations lend credence to this model also (Schwartz and Coppersmith, 1984; Sieh 1978a,b 1981). The recurrence curve might be written:

$$n(M) = 10^{a-bM} H(M^a - M) + \delta(M_{\max} - M) \quad (3.5)$$

There are a few observations from single long fault zones (Case II). Lahr and Stevens (1982) suggested that the distribution curve in Alaska deviates to a smaller b-value at large magnitudes. Singh et al. (1983) have made a similar observation for the Mexican subduction zone.

Anderson and Bodin (1986) have suggested that a distribution of rupture lengths for large earthquakes can be found on the Cerro Prieto

and Imperial fault system (Case II), with multiple rupture length large earthquakes causing failure of the same point in the fault. They consider a plausible explanation to be that barriers which might limit rupture length are distributed randomly along the fault. This implies an exponential distribution of rupture length

$$P(L) = P_0 \exp(-\beta L) \quad (3.6)$$

and a double-exponential magnitude distribution for large earthquakes. Like Wesnousky et al. (1983), Anderson and Bodin (1986) would retain the exponential distribution for small magnitudes. They specifically propose that the transition in behavior should occur at the transition from small to large earthquakes. Based on this model, one might write the magnitude distribution as:

$$N(M) = A(e^{\overline{b}(M_t - M)} - 1)H(M_t - M) + B[e^{-\beta L(M)} - e^{-\beta L(M_{\max})}]H(M_{\max} - M) \quad (3.7)$$

where M_t is the magnitude of the transition from small to large, plate-rupturing earthquakes. The function $L(M)$ gives rupture length of large earthquakes as a function of magnitude. The transition from small to large, plate-rupturing earthquakes occurs where $L(M_t)$ approximately equals the width of the seismogenic zone.

Main and Burton (1984) have employed a theorem from information theory to derive the density distribution function:

$$n(M) = C \exp(-b_1 M - b_2 M_0(M)) . \quad (3.8)$$

They also show a procedure to derive b_1 and b_2 from the magnitude distribution at small magnitude and the moment rate. We observe that b_2 will typically take a value of $(M_0(M))^{-1}$ for some value of M which is a little smaller than $M_{\max}$. Therefore, since $M_0(M)$ changes very rapidly, at small magnitudes the term in b_2 is order of magnitudes smaller than the term in b_1 , and equation (3.8) decays exponentially in the usually assumed manner. When M_0 gets large, the term in b_2 quickly becomes dominant and causes $n(M)$ to drop more rapidly. Thus equation (3.8) is comparable to $n_3(M)$ in Anderson and Luco (1983). An excess of large magnitude earthquakes, the concern of the investigations mentioned in the preceding paragraphs, cannot be modeled by equation (3.8).

THE PARAMETER $M_{\max}$

The largest earthquake to be used in the earthquake occurrence rate curve is designated as $M_{\max}$, corresponding to moment $M_{o,\max}$. A complete model for $M_{o,\max}$ has not been presented. Before discussing existing models, it is worthwhile to make a few general comments about properties which $M_{o,\max}$ must obey. The first is that $M_{o,\max}$ must be greater than or equal to the magnitudes of all the large historical earthquakes in the seismic zone:

$$M_{o,\max} \geq k M_o \quad \text{all } k. \quad (4.1)$$

On the other hand, an upper bound can be derived from the dimension of the fault or diffuse zone. For a fault with dimensions L_z and W_z , an extreme estimate for the moment that fault is capable of supporting is:

$$M_{o,\max} \geq \mu L_z W_z u_z \quad (4.2)$$

in which u_z is related to L_z and W_z by the extreme values of stress drop which are possible. Scholz (1982) has suggested the following averages:

$$\begin{aligned} u_z &= 1.25 \times 10^{-5} L_z \quad \text{for strike slip faults} \\ u_z &= 2 \times 10^{-5} L_z \quad \text{for thrust faults.} \end{aligned} \quad (4.3)$$

Thus a value of $u_z = 4 \times 10^{-5} L_z$ is a conservative bound, but the average values in equation (4.3) are probably adequate when one considers that L_z and W_z are physical upper bounds which may not be realized in most cases. For a volume undergoing more or less uniform strain, L_z and W_z can be taken from the dimensions of the zone.

In the western United States, the maximum earthquake is a large earthquake in essentially all cases. For a diffuse zone with no conspicuous faults, as in much of the eastern United States, the failure to recognize surface faulting could possibly imply that there are no large earthquakes occurring in the region. (The New Madrid, Charleston, and La Malbie regions are likely excepted from this discussion.) If that hypothesis were to receive experimental or theoretical support, the upper limit on the moment of an earthquake which is constrained to be small in the sense of Scholz (1982) could be used to apply a more restrictive estimate for $M_{0,max}$. This hypothesis thus deserves added research to test its validity. Most of the largest earthquakes which have been catalogued in the eastern United States are probably small in this sense. The extreme moment in such a case would be on the order of $M_0^u = (3.6 \times 10^{11} \text{ dyne-cm}^{-2})(15 \times 10^5 \text{ cm})^3(2 \times 10^{-5}) = 2.4 \times 10^{25} \text{ dyne-cm}$. This corresponds, by the moment-magnitude relation of Hasegawa (1983), to a magnitude of 6.4, and by the moment-magnitude relation of Nuttli (1983), to a magnitude of 6.1.

Models for Determining M_{max}

Fault Dimension. M_{max} can be determined from a combination of theoretical and empirical relations, or from purely empirical relations. The simplest procedures are based on the fault dimensions. The extreme moment, $M_{0,max}$, was given above in equation (4.2). Wesnousky et al. (1983) have employed these relations to calculate the maximum moment of earthquakes for faults in Japan. From the extreme moment, an extreme magnitude is established by means of equation (2.3).

An alternative approach, which has been used much more extensively, is to use an entirely empirical approach based on correlations of length of rupture and magnitude (e.g. Slemmons, 1977), in combination with an empirical estimate of the extreme fraction of the total fault length which is likely to rupture in a single earthquake. Conceptually, at least, this approach should give the same result as the approach based on extreme moment, provided the extreme fraction of fault length is taken to be 1.0.

Smith (1976) and Papastamatiou (1980) have used a very different approach of using moment rate constraints to estimate $M_{\max}$. Papastamatiou assumed that the occurrence rate at small magnitude is stationary, and the shape of the occurrence rate curve at large magnitude was known, and thus chose $M_{\max}$ where the occurrence rate curve needed to be cut off to balance the moment rate. Smith (1976) assumed that at least one maximum magnitude event was responsible for observed Holocene deformation (i.e. he assumed an upper bound for the recurrence time of the maximum event) and was thus able to constrain $M_{\max}$. Some of the mathematical procedures for obtaining $M_{\max}$ in this manner are elaborated in Anderson and Luco (1983).

McGuire (1977) has derived a probability distribution for $M_{\max}$ based on an assumed shape for $N(M)$ and the probability of seeing large earthquakes based on the duration of the observational period.

There is a class of models, all referred to as characteristic earthquake models, which assumes that all large earthquakes on a given plate margin are more or less the same size (e.g. Schwarz and Coppersmith, 1984; Sykes and Nishenko, 1984). Thus under this model, if a large

earthquake has occurred on a fault, it is assumed that future large earthquakes will more or less repeat at this particular size. This model is convenient in allowing one to make predictions, but the data to support this model is still rather tenuous.

Lay et al. (1982) have developed an interesting approach for estimating $M_{\max}$ on convergent plate margins. They observed a regularity in the maximum historical magnitude when plotted on axes of age of the subducted ocean floor and rate of convergence. It is possible that an analogous approach could be developed for the intraplate areas such as the eastern United States. Regions which are analogous to the stable continental margin of eastern North America might be the stable margins of the Arctic, South America, Europe, India, or Australia. Parameters which could be studied would include the age of the margin and spreading rate at the mid-ocean ridge. Features which are analogous to the Appalachian or Ouchita Mountains could include other ancient mountains such as the Urals of eastern Europe, Caledonides of England, Kjolén Mountains of Norway, the Great Dividing Range of Australia, the Ghat Ranges of India, and the various ranges of south and west Africa and Madagascar. Features comparable to the stable interior are interior areas of South America, Africa, western Europe, Australia, and India. Research on this approach has a reasonable chance of yielding useful results.

CONDITIONAL PROBABILITIES

Along major plate margins, it is observed that the rupture areas of the largest earthquakes are approximately delineated by the extent of the aftershock zone. When multiple large earthquakes are considered along a specific plate margin, aftershock zones abut with little overlap. Repeat ruptures of a specific segment are separated by several decades in most cases. This has led to the seismic gap theory. Sykes (1983) presents a review of this theory. Critical papers in the development of the theory include Tobin and Sykes (1968), Sykes (1971), Davies, et al. (1981), Kelleher, et al. (1973), McCann, et al. (1979), Nishenko and McCann (1981), and Ruff and Kanamori (1980,1983). The field will not be reviewed in detail here.

A corollary of the concept of seismic gaps is that there should be some regularity in the occurrence rates of the largest earthquake along any particular portion of a plate boundary. Thus some statistical models have been devised to quantify earthquake probabilities within specific gaps.

There are, of course, many methods to quantify these observations. The essence of this quantification is to apply conditional probabilities: given some information on recurrence rates of deformation rates and the time since the last earthquake, a conditional probability can be derived which in general differs from the probability in a time interval chosen entirely at random. Papers which have pursued these methods include Veneziano and Cornell (1974), Kameda and Ozaki (1979), Patwardhan, et al. (1980), Savy, et al. (1980), Kameda and Takagi (1981), Lindh (1983), Jacob (1984), Sykes and Nishenko (1984), and

Kiremidjian and Anagnos (1984). The question is whether conditional probabilities might be useful to qualify the earthquake occurrence rates, as obtained from slip rate or strain rate constraints, in the eastern United States.

It is assumed that the recurrence times between earthquakes is a random function obeying a probability distribution $p(t)$. For example, a Poisson probability density for recurrence time would be written as:

$$p(t) = \nu \exp(-\nu t) \quad (5.1)$$

where ν^{-1} is the mean recurrence time. Then define:

$$P_I(t_1, t_2) = \int_{t_1}^{t_2} p(t) dt \quad (5.2)$$

as the probability of a particular earthquake occurring in the time interval between t_1 and t_2 , where $t = 0$ refers now to the time of the last earthquake in this location. Now suppose T years have occurred since the last earthquake, and we desire the probability P_0 of the next earthquake occurring within the next Δ years. This result is simply (Kameda and Ozaki, 1979):

$$P_\Delta = \frac{P_I(T, T+\Delta)}{P_I(T, \infty)} \quad (5.3)$$

In probabilistic seismic risk analysis, it is common to employ a Poisson distribution (5.1) for the occurrence times of earthquake main shocks (e.g., Cornell, 1968; Kameda and Ozaki, 1979). Under this assumption, P_Δ is independent of T . Kameda and Ozaki (1979) found that the Poisson assumption was not supported by the earthquake record at Kyoto, Japan.

They modeled the Kyoto data by a double Poisson process, which they defined as a Poisson process with one (or sometimes two) step increases in the Poisson rate parameter as the time T increases. Kameda and Ozaki also found evidence that longer waiting times were followed by larger earthquakes, and presented a model in which the size of the earthquake was also conditional on T .

Kameda and Takagi (1981) employed a somewhat more sophisticated model. They divided earthquakes in the Kinki district of Japan (including Kyoto) into the interplate earthquakes (offshore, caused by subduction of the Philippine Sea plate below the Asian plate) and intraplate earthquakes (caused by deformation of the island of Japan). Recurrence times for the interplate earthquakes were modeled by a truncated Poisson distribution.

$$P_I(t, \infty) = \begin{cases} 0 & t < t_0 \\ \frac{1 - \exp[-v(t - t_0)]}{1 - \exp[-v(t_u - t_0)]} & t_0 \leq t \leq t_u \\ 1 & t_u < t \end{cases} \quad (5.4)$$

The truncation causes P_Δ to increase with time.

Kameda and Takagi modeled the intraplate earthquakes with a Poisson model which was generalized from Kameda and Ozaki (1979). They divided the time between successive interplate earthquakes into four intervals. In each interval, the rate parameter for the Poisson process was given a different value. Specifically, the rate is high for a short period immediately following the great plate rupturing earthquake. It then drops to zero, and increases by steps until the next great earthquake.

Savy, et al. (1980) also worked with the concept of a Poisson process in which the rate is an increasing function of time. When $v(t)$ is a power law, the recurrence times obey a Weibull distribution:

$$\begin{aligned} P(t) &= 1 - \exp(-\xi t^\eta) \\ p(t) &= \xi t^{\eta-1} \exp(-\xi t^\eta) \end{aligned} \quad (5.5)$$

which Savy, et al. employed in their risk analysis. The paper also introduced the instantaneous return period, equal to the inverse of the failure rate function:

$$T_i(t) = [h(t)]^{-1} = [1 - P(t)]/p(t) \quad (5.6)$$

Savy, et al. estimated that the parameters of the Weibull distribution appropriate for the San Andreas fault as represented at Pallett Creek are $\xi = 1.326 \times 10^{-11}$ and $\eta = 4.446$.

Lindh (1983) evaluated equation (5.3) for several faults in California. For his study, a Gaussian probability density for recurrence time was assumed:

$$p(t) = \frac{1}{\sigma_R \sqrt{2\pi}} \exp\left[-\frac{(t - T_R)^2}{2\sigma_R^2}\right] \quad (5.7)$$

In this equation, T_R is the mean recurrence time, and σ_R is the standard deviation of the recurrence time. Compared to the Weibull distribution, equation (5.5), a Gaussian model offers the conceptual disadvantage that negative recurrence times are not formally excluded. In practice, the tail of equation (5.7) at small and negative time is not important, and the Gaussian model offers the advantage that the parameters T_R and

and σ_R are more easily and intuitively associated with properties of the distribution than are ξ and η in equation (5.5). We note that neither distribution has been adequately tested by data or suggested by theoretical arguments. By assumption, Lindh uses $\sigma_R = 0.3T_R$. The mean recurrence intervals, likewise, were not well known in many cases, but the paper is probably successful in ranking the faults which were considered by relative probabilities of earthquake occurrence.

The paper by Jacob (1984) compared (5.7) and a lognormal distribution:

$$p(t) = \frac{1}{\sigma_L \sqrt{2\pi}} \exp\left[-\frac{(\log t - \log T_L)^2}{2\sigma_L^2}\right] \quad (5.8)$$

For gaps, they show that probabilities based on (5.7) exceed those derived from distribution (5.8). Equation (5.8) fits the recurrence time distribution in the Aleutian arc slightly better than equation (5.7).

Sykes and Nishenko (1984) employed equation (5.7) exclusively, and obtained results which are consistent with Lindh (1983) for a similar selection of faults. Implicit in these articles is an assumption that specific fault segments tend to rupture repeatedly with about the same amount of slip in each earthquake. (This is referred to as "characteristic earthquakes" by Sykes and Nishenko; other authors have defined characteristic earthquakes in different ways.)

The paper by Patwardhan, et al. (1980) develops a semi-Markov model for recurrence of great earthquakes. This model is motivated by plate tectonic theory of seismic gaps and recognized a probabilistic relationship between the times between earthquakes and the magnitude

of these earthquakes. The primary result is a set of probabilities of occurrence of different magnitudes of earthquakes in a given zone during a specific period of interest. Compared with the previous model (equations (5.2) and (5.3)), the additional effort to put probabilities on the magnitudes of the successive earthquakes greatly increases the complexity of the model.

Kiremidjian and Anagnos (1984) considered a probabilistic model with the realistic property of a seismic process that depends on what has happened in the past. In their model, the waiting times between large earthquakes are represented by the Weibull distribution (equation (5.5)). The slip of an earthquake is represented as a function of the holding time since the previous earthquake follows the slip-predictable model of Shimazaki and Nakata (1980). The introduction of the slip predictable model to calculate slip, and thus magnitude, represents a quantitative framework for the observations of Kameda and Ozaki (1979) and a possible conceptual improvement over the other models described above, although the slip-predictable model has not yet been adequately verified.

For the eastern United States, the use of models which employ conditional probabilities would possibly have a major impact on seismic risk analysis. For example, if in the Charleston region the recurrence time of earthquakes may be several hundred years, the interval since the last earthquake of only one hundred years would imply, by these models, that the probability of a repeat event is very low at this time. If, however, other localities along the eastern seaboard have the potential to cause a significant earthquake, then the holding time

at each of these would be unknown but potentially large, with the result that the hazard at these localities would be enhanced. The papers of McGuire (1979) and McGuire and Bernhard (1981) are relevant to this discussion for the documentation of how well the immediate past serves as a prediction of future seismic hazard in China. Based on these papers, it would seem that the use of conditional probabilities in the eastern United States is not justified.

SEISMIC STRAIN RATES IN THE EASTERN UNITED STATES

Anderson (1986) presents an estimate of seismic strain rates in the eastern United States. The rates are summarized in Figure 6.1. Anderson (1986) discusses the procedure which was employed to obtain the model, the problem of verifying the model, and the observation that in many regions, denudation stresses are sufficient to account for the observed rates. Highlights of his findings are the following:

1. Strain rate implications of a seismicity model should be considered when the model is prepared.
2. In most of the eastern United States, strain rates are one to two orders of magnitude too small to be observed from the surface observations because of denudation.
3. It is possible that denudation stresses trigger some of the earthquakes.

6.1 Implications of Denudation and Sedimentation Rate Hypothesis

Where the denudation and sedimentation rate hypotheses of Anderson (1986) are true, one can use them to quantify seismic risk in the following ways.

1. The implication for tectonic style is that EUS earthquakes are without tectonic significance for deformation of the crust.
2. The maximum magnitude earthquake would be expected to be confined entirely to the crust, and large, plate-rupturing earthquakes would not occur. This provides a physical basis for estimating $M_{\max}$ and for choosing an exponential distribution for earthquake magnitudes.

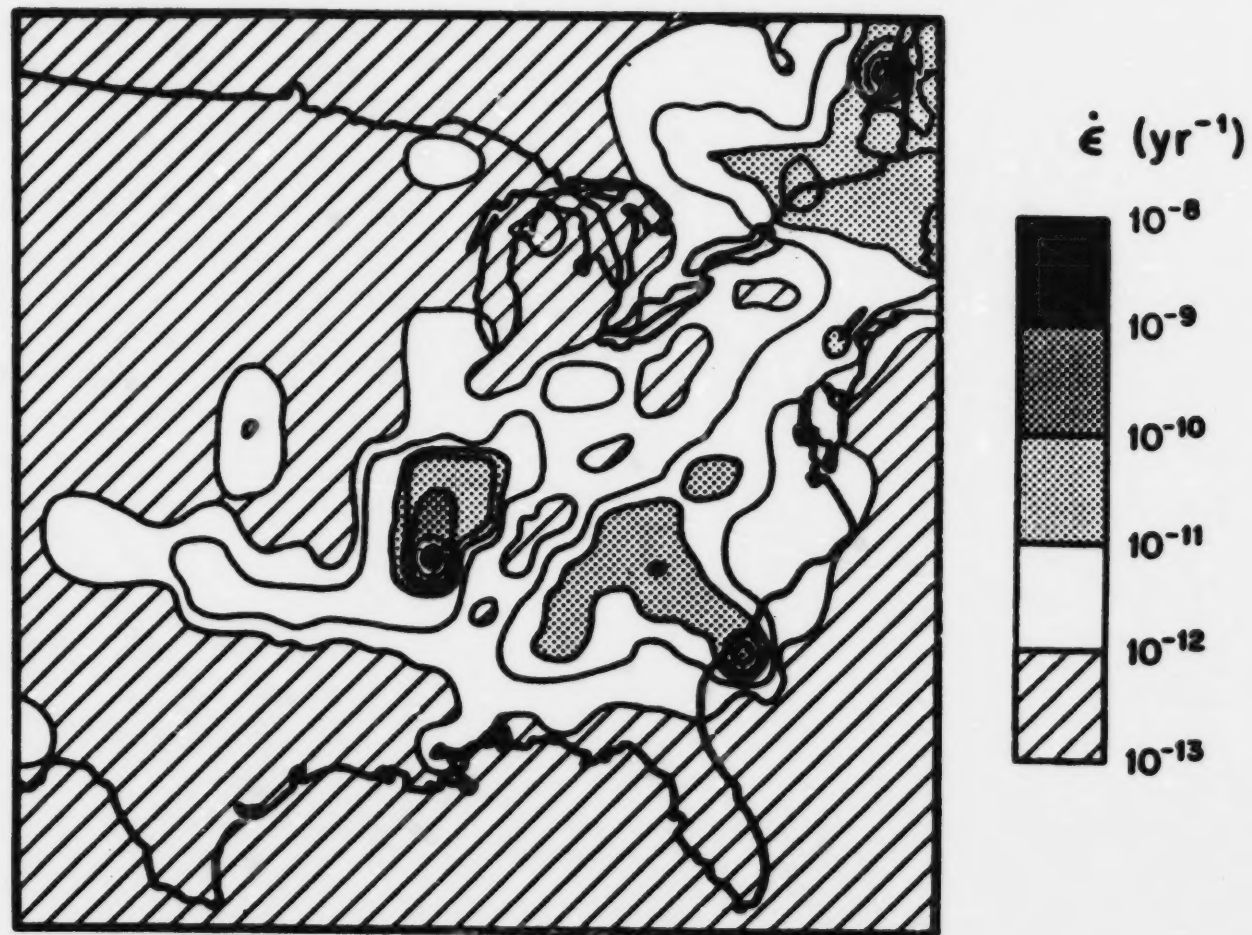


Figure 6.1. Preliminary map of seismic strain rates in the Eastern United States.

3. Denudation and sedimentation rates can be measured everywhere and employed to place a lower bound on seismic moment rates.

ACKNOWLEDGEMENTS

This report has been prepared by the author upon the request of Structural and Earthquake Engineering Consultants, Inc., with the sponsorship of the United States Nuclear Regulatory Commission.

REFERENCES

- Aki, K. (1966). Generation and propagation of G-waves from the Niigata earthquake of June 16, 1964, 2. Estimation of earthquake moment, released energy, and stress strain drop from the G-wave spectrum, Bull. Earthquake Research Inst., Tokyo University 44, 73-88.
- Aki, K. (1967). Scaling law of seismic spectrum, J. Geophys. Res., 72, 1217-1232.
- Anderson, J. G. (1984). U.S.G.S. Open File Report, Menlo Park, California.
- Anderson, J. G. and P. Bodin (1986). Earthquake recurrence models and historical seismicity in the Mexicali - Imperial Valley, submitted for publication.
- Anderson, J. G. and J. E. Luco (1983). Consequence of slip rate constraints on earthquake occurrence relations, Bull. Seism. Soc. Amer., 73, 471-496.
- Anderson, J. G. (1979). Estimating the seismicity from geological structure for seismic-risk studies, Bull. Seism. Soc. Amer., 69, 135-158.
- Anderson, J. G. (1986). Seismic strain rates in the central and eastern United States, Bull. Seism. Soc. Amer., (in press).
- Anderson, J. G. (1984). Synthesis of seismicity and geological data in California, U.S. Geological Survey, Open-File Report, 84-0424, Boulder, CO.
- Andrews, D. J. (1980). A stochastic fault model, 1. static case, J. Geophys. Res., 85, 3867-3877.
- Ben-Menahem, A. S. J. Singh and F. Solomon (1969). Static deformation of a spherical earth model by internal dislocations, Bull. Seism. Soc. Amer., 59, 813-854.
- Brune, J. N. (1968). Seismic moment, seismicity, and rate of slip along major fault zones. J. Geophys. Res., 73, 777-784.
- Brune, J. N. and C. R. Allen (1967). A low-stress-drop, low magnitude earthquake with surface faulting: the Imperial, California earthquake of March 4, 1966, BSSA 57, 501-
- Brune J. N., A. Espinosa and J. Oliver (1963). Relative excitation of surface waves by earthquakes and underground explosions in the California-Nevada region, J. Geophys. Res., 68, 3501-3513.

- Brune, J. N. and C.-Y. King (1967). Excitation of mantle Rayleigh waves of period 100 seconds as function of magnitude, BSSA, 57, 1355-1365.
- Bufe, C. G., P. W. Harsh, and R. O. Burford (1977). Steady-state seismic slip -- A precise recurrence model, Geophys. Res. Letts., 4, 91-94.
- Caputo, M. and R. Console (1980). Statistical distribution of stress drops and faults of seismic regions. Tectonophysics, 67, 713-720.
- Chen, W. P. and P. Molnar (1977). Seismic moments of major earthquakes and the average rate of slip in Central Asia, J. Geophys. Res., 82, 2945-2969.
- Chen, W. P. and P. Molnar (1983). Focal depths of intracontinental and intraplate earthquakes and their implications for the thermal and mechanical properties of the lithosphere, J. Geophys. Res., 88, 4183-4214.
- Cornell, C. A. (1968). Engineering seismic risk analysis, Bull. Seism. Soc. Amer., 58, 1583-1606.
- Counselmann, C. C., III (1983). Very long baseline interferometric geodesy with GPS satellites, Summaries of Technical Reports XVI, OFR 83-525, 174-176.
- Davies, J. N., L. R. Sykes, L. House, and K. H. Jacob (1981). Shumagin seismic gap, Alaska peninsula: history of great earthquakes, tectonic setting, and evidence for high seismic potential. J. Geophys. Res., 86, 3821.
- Davies, G. F. and J. M. Brune (1971). Regional and global fault slip rates from seismicity, Nature Physical Science, 229, 101-107.
- Deng, Q., Y. Change, K. Hsu and F. Fan (1971). On the tectonic stress field in China and its relation to plate movement. Phys. Earth and Planetary Interiors, 18, 257-273.
- Doser, D. I. and R. B. Smith (1982). Seismic moment rates in the Utah region, Bull. Seism. Soc. Amer., 72, 525-551.
- Doser, D. I. and R. B. Smith (1983). Seismicity of the Teton - southern Yellowstone region, Wyoming, BSSA, 73, 1369-1394.
- Greensfelder, R. W., F. C. Kintzer, and M. R. Somerville (1980). Seismo-tectonic regionization of the Great Basin, and comparison of moment rates computed from Holocene strain and historic seismicity, in Proceedings of Conference X, Earthquake Hazards Along the Wasatch and Sierra-Nevada Frontal Fault Zones, Open-File Rept. 80-801, United States Department of the Interior Geological Survey, Menlo Par, California, 433-493.

- Greensfelder, R. W., F. C. Kintzer, and M. R. Somerville (1980). Seismotectonic regionalization of the Great Basin, and comparison of moment rates computed from Holocene strain and historic seismicity: summary, Bull. Geo. Soc. Amer., 97, 518-123.
- Gutenberg, B. and C. F. Richter (1954). Seismicity of the Earth and Associated Phenomena, 2nd Edition, Princeton Univ. Press, Princeton New Jersey, 310.
- Hanks, T. C. and H. Kanamori (1979). A moment magnitude scale, J. Geophys. Res., 84, 2348-2350.
- Hanks, T. C. and D. M. Boore (1984). Moment-magnitude relations in theory and practice, J. Geophys. Res., 89, 6229-6235.
- Hasegawa, H. S. (1983). Lg spectra of local earthquakes recorded by the eastern Canada telemetered network and spectral scaling, Bull. Seism. Soc. Amer., 73, 1041-1061.
- Heaton, T. H. and H. Kanamori (1984). Seismic potential associated with subduction in the northwestern United States, Bull. Seism. Soc. Amer., 74, 933-941.
- Hyndman, R. D. and D. H. Weichert (1983). Seismicity and rates of relative motion on the plate boundaries of western North America, Geophysical Journal of the Royal Astronomical Society, 72, 59-82.
- Jacob, K. H. (1984). Estimates of long-term probabilities for future great earthquakes in the Aleutians, Geophys. Res. Letters, 11, 295-298.
- Kameda, H. and H. Takagi (1981). Seismic hazard estimation based on non-Poisson earthquake occurrences, Memoirs of the Faculty of Engineering, Kyoto University, Vol. XLIII, 397-433.
- Kameda, H. and Y. Ozaki (1979). A renewal process model for use in seismic risk analysis, Memoirs of the Faculty of Engineering, Kyoto University, Vol. XLI, 11-35.
- Kelleher, J. (1972). Rupture zones of large South American earthquakes and some predictions, J. Geophys. Res., 77, 2087.
- Kelleher, J., L. R. Sykes and J. Oliver (1973). Possible criteria for predicting earthquake locations and their application to major plate boundaries of the Pacific and the Caribbean, J. Geophys. Res., 78, 2547.
- Kiremidjian, A. S. and T. Anagnos (1984). Stochastic slip-predictable model for earthquake occurrences, BSSA, 74, 739-755.
- Kostrov, V. V. (1974). Seismic moment and energy of earthquakes, and seismic flow of rock, IZV. Acad. Sci., USSR Phys. Solid Earth, 1, 23-44.
- Lahr, J. C. and C. D. Stevens (1982). Alaska seismic zone: possible example of non-linear magnitude distribution for faults, abstract, Earthquake Notes, 53, No. 1, 66.

- Lay, T., H. Kanamori and L. Ruff (1982). The asperity model and the nature of large subduction zone earthquakes, Earthquake Prediction Research, 1, 3-71.
- Leblanc, G. (1983). A basis for comparing LaMalbaie, Quebec to Charleston, South Carolina, in Proceedings of Conference XX, A Workshop on "the 1886 Charleston, South Carolina, earthquake and its implications for today," W. W. Hays and P. L. Gori, editors, Reston Virginia, 435-455, U.S.G.S. Open File Report 83-843.
- Lindh (1983). Preliminary assessment of long-term probabilities for large earthquakes along selected fault segments of the San Andreas fault system in California, U.S.G.S. Open-File Report 83-63, Menlo Park, California.
- Lyon-Caen, H. and P. Molnar (1983). Constraints on the structure of the Himalaya from an analysis of gravity anomalies and a flexural model of the lithosphere, J. Geophys. Res., 88, 8171-8191.
- Main, I. G. and P. W. Burton (1981). Rates of crustal deformation inferred from seismic moment and Gumbel's third distribution of extreme magnitude values, in Earthquakes and Earthquake Engineering -- Eastern United States, J. E. Beavers, Editor, Ann Arbor Science Publishers, Inc., Ann Arbor, Michigan, 937-951.
- Main, I. G. and P. W. Burton (1984). Information theory and the earthquake frequency - magnitude distribution, BSSA, 74, 1409-1426.
- McCann, W. R., S. P. Nishenko, L. R. Sykes and J. Krause (1979). Pure Appl. Geophys., 117, 1082.
- McGuire, R. K. (1977). Effects of uncertainty in seismicity on estimates of seismic hazard for the east coast of the United States, BSSA, 67, 827-848.
- McGuire, R. K. (1979). Adequacy of simple probability models for calculating felt-shaking hazard, using the Chinese earthquake catalog, Bull. Seism. Soc. Amer., 69, 877-892.
- McGuire, R. K. and T. P. Bernhard (1981). Effects of temporal variations in seismicity on seismic hazard, Bull. Seism. Soc. Amer., 71, 321-334.
- McNally, K. C. and J. B. Minster (1981). Nonuniform seismic slip rates along the Middle America Trench, Journal of Geophysical Research, 86, 4949-4959.
- Molnar, P. (1979). Earthquake recurrence intervals and plate tectonics, Bull. Seism. Soc. Amer., 69, 115-133.
- Molnar, P. (1983). Average regional strain due to slip on numerous faults of different orientations, J. Geophys. Res., 88, 6430-6432.

- Molnar, P. and P. Tapponnier (1975). Cenozoic tectonics of Asia: effects of a continental collision, Science, 189, 419-425.
- Molnar, P. and P. Tapponnier (1978). Active tectonics of Tibet, J. Geophys. Res., 83, 5361-5375.
- Molnar, P. and W. P. Chen (1982). Seismicity and mountain building, in Mountain Building Processes, edited by V. Briegel and K. Hsu, 41-57, Academic, New York.
- Molnar, P. and W. P. Chen (1983). Focal depths and fault plane solutions of earthquakes under the Tibetan Plateau, J. Geophys. Res., 88, 1180-1196.
- Molnar, P. and Deng Qidong (1984). Faulting associated with large earthquakes and the average rate of deformation in central and eastern Asia, J. Geophys. Res., 89, 6203-6227.
- Nishenko, S. P. and W. R. McCann (1981). In Earthquake Prediction, An International Review, Maurice Ewing Series 4, edited by D. W. Simpson and P. G. Richards, Washington D.C., 20.
- Nuttli, O. W. (1983). Average seismic source parameters for mid-plate earthquakes, Bull. Seism. Soc. Amer., 73, 519-535.
- Papastamatiou, D. (1980). Incorporation of crustal deformation to seismic hazard analysis, Bull. Seism. Soc. Amer., 70, 1321-1335.
- Patwardhan, A. S., R. B. Kulkarni, and D. Tocher (1980). A Semi-Markov model for characterizing recurrence of great earthquakes, Bull. Seism. Soc. Amer., 70, 323-347.
- Peterson, E. T. and T. Seno (1984). Factors affecting seismic moment release rates in subduction zones, J. Geophys. Res., 89, 10,233-10,248.
- Reichle, M. and I. Reid (1977). Detailed study of earthquake swarms from the Gulf of California, BSSA, 67, 159-171.
- Reichle, M. G. F. Sharman and J. N. Brune (1976). Sonobuoy and teleseismic study of Gulf of California transform fault earthquake sequences, BSSA, 66, 1623-1641.
- Reid, H. F. (1910). The mechanism of the earthquake, the California earthquake of April 18, 1906, report of the State Earthquake Investigation Commission, Vol. 2, Carnegie Institution of Washington D.C., 192.
- Ruff, L. and H. Kanamori (1980). Seismicity and the subduction process. Phys. Earth Planet Interiors, 23, 240.

- Ruff, L. and H. Kanamori (1983). The rupture process and asperity distribution of 3 great earthquakes from long period diffracted T-waves, Phys. Earth. Planet Interiors, Vol. 31, 202-230.
- Savy, J. B., H. C. Shah, and D. Boore (1980). Nonstationary risk model with geophysical input, J. Struc. Div., ASCE, 106, 145-163.
- Scholz, C. H. (1982). Scaling laws for large earthquakes: consequences for physical models, Bull. Seism. Soc. Amer., 72, 1-14.
- Schwartz, D. P. and K. J. Coppersmith (1984). Fault behavior and characteristic earthquakes: examples from the Wasatch and San Andreas fault zones, J. Geophys. Res., 89, 5681-5698.
- Shimazaki, K. and T. Nakata (1980). Time predictable recurrence for large earthquakes, Geophys. Res. Letts., 7, 279-282.
- Sieh, K. E. (1978a). Prehistoric large earthquake produced by slip on the San Andreas fault at Pallett Creek, California, J. Geophys. Res., 83, 3907-3939.
- Sieh, K. E. (1978b). Slip along the San Andreas fault associated with the great 1857 earthquake, Bull. Seism. Soc. Amer., 68, 1421-1448.
- Sieh, K. E. (1981). A review of geological evidence for recurrence times of large earthquakes, in Earthquake Prediction, an International Review, D. W. Simpson and P. G. Richards, editors, American Geophysical Union, Washington D.C., 181-207.
- Singh, S. K., L. Astiz and J. Havskov (1981). Seismic gaps and recurrence periods of large earthquakes along the Mexican subduction zone, Bull. Seism. Soc. Amer., 71, 827-843.
- Singh, S. K., M. Rodriguez and L. Esteva (1983). Statistics of small earthquakes and frequency of occurrence of large earthquakes along the Mexican subduction zone, BSSA, 73, 1779-1796.
- Slemmons, D. B. (1977). Faults and earthquake magnitude, in "State-of-the-art for assessing earthquake hazards in the United States," U.S. Army Engineers Waterways Experiment Station Misc. Paper S-73-1, Report 6, Vicksburg, Mississippi.
- Smith, S. W. (1976). Determination of maximum earthquake magnitude, Geophys. Res. Letts., 33, 351-354.
- Sykes, L. R. (1971). Aftershock zones of great earthquakes, seismicity gaps, and earthquake prediction for Alaska and the Aleutians, J. Geophys. Res., 76, 8021.
- Sykes, L. R. (1983). Predicting great earthquakes, in Earthquakes: Observation, Theory and Interpretation, Soc. Italiana de Fisica, Bologna, Italy.

- Sykes, L. R. and R. C. Quittmeyer (1981). Repeat times of great earthquakes along simple plate boundaries, in Earthquake Prediction, an International Review, D. W. Simpson and P. G. Richards, editors, American Geophysical Union, Washington D.C., 217-247.
- Sykes and Nishenko (1984). Probabilities of occurrence of large, plate rupturing earthquakes for the San Andreas, San Jacinto and Imperial faults, California, J. Geophys. Res., 89, 5905-5927.
- Talwani, P. (1983). Tectonic Models - Old and New, in U.S.G.S. Open File 83-843, Proc. of Conf. XX, 88-89.
- Tapponnier, P. and P. Molnar (1976). Slip line field theory and large scale continental tectonics, Nature, 264, 319-324.
- Tapponnier, P. and P. Molnar (1977). Active faulting and Cenozoic tectonics of China, J. Geophys. Res., 82, 2905-2930.
- Tapponnier, P. and P. Molnar (1979). Active faulting and Cenozoic tectonics of the Tien Shan, Mongolia and Baykal region, J. Geophys. Res., 84, 3425-3459.
- Tobin, D. and L. R. Sykes (1968). Seismicity and tectonics of the northeast Pacific ocean, J. Geophys. Res., 73, 3821.
- Veneziano, D. and C. A. Cornell (1974). Earthquake models with spatial and temporal memory for engineering seismic risk analysis, Department of Civil Engineering, Massachusetts Institute of Technology, R74-18, Cambridge, Massachusetts.
- Wang, S. C., K. C. McNally and R. J. Geller (1982). Seismic strain release along the Middle American Trench, Mexico, Geophys. Res. Letts., 9, 182-185.
- Wallace, R. E. (1970). Earthquake recurrence intervals on the San Andreas fault, Bull. Geol. Soc. Amer., 81, 2875-2890.
- Weichert, D. H. and R. D. Hyndman (1983). A comparison of the rate of seismic activity and several estimates of deformation in the Puget Sound area, in J. C. Yount and R. S. Crosson, editors, Proceedings of Workshop XIV, Earthquake Hazards of the Puget Sound Region, Washington, Open File Report, 83-19, U.S. Geological Survey, Menlo Park, California, 105-130.
- Wentworth, C. M. (1983). Reverse faulting as the source of earthquakes along the eastern seaboard of the United States: problems and needed research in U.S.G.S. Open File Report 83-843, Reston, Virginia, Proc. of Conf. XX, A Workshop on "the 1886 Charleston, South Carolina, earthquake and its implications for today," W. W. Hays and P. L. Gori, editors, 117-125.

- Wesnousky, S. G., C. H. Scholz, and K. Shimazaki (1982). Deformation of an island arc: rates of moment release and crustal shortening in intraplate Japan determined from seismicity and quaternary fault data, J. Geophys. Res., 87, 6829-6852.
- Wesnousky, S. G., C. H. Scholz, K. Shimazaki, and T. Matsuda (1983). Earthquake frequency distribution and the mechanics of faulting, J. Geophys. Res., 88, 9334-9340.
- Wyss, M. and J. N. Brune (1968). Seismic moment, stress, and source dimensions for earthquakes in the California-Nevada region, J. Geophys. Res., 73, 4681-4694.

APPENDIX II

UNIFORM RISK SPECTRA OF STRONG
EARTHQUAKE GROUND MOTION:
NEQRISK:

by

V. W. Lee and M. D. Trifunac

ABSTRACT

The concept of uniform risk spectra of Anderson and Trifunac (1977) has been generalized to include (1) more refined description of earthquake source zones, (2) the uncertainties in estimating seismicity parameters a and b in $\log_{10} N = a - bM$, (3) to consider uncertainties in estimation of maximum earthquake size in each source zone, and to (4) include the most recent results on empirical scaling of strong motion amplitudes at a site.

Examples of using to new NEQRISK program are presented and compared with the corresponding case studies of Anderson and Trifunac (1977). The organization of the computer program NEQRISK is also briefly described.

CONTENTS

ABSTRACT.....	II. i
1. INTRODUCTION.....	II. 1
2. MODEL OF SEISMICITY.....	II. 4
2.1 Types of Faults.....	II. 4
2.2 Estimation of Seismicity.....	II. 5
2.3 The Source Elements.....	II. 7
3. ATTENUATION OF RISK FUNCTIONALS FROM THE SEISMIC SOURCES.....	II. 12
3.1 Previous Work	II. 12
3.2 The New Scaling Functions.....	II. 15
3.3 Comparison of the Old and New Scaling Functions.....	II. 18
3.4 Derivation of the Function $q_{ij}(S(\omega))$	II. 23
4. MODEL OF SEISMIC RISK.....	II. 27
4.1 The Uniform Risk Spectra.....	II. 27
4.2 Formulation of the Risk Model.....	II. 28
4.3 Characterizing Uncertainties of Seismicity.....	II. 35
4.4 The Probability of Exceedance $P[S(\omega)]$	II. 41
4.5 Literal Seismicity.....	II. 46
4.6 Summary.....	II. 52
5. THE COMPUTER PROGRAM.....	II. 53
5.1 Introduction.....	II. 53
5.2 Input Data.....	II. 56
5.3 The Flow Chart and Program Description.....	II. 64
5.4 Listing of Sample Inputs.....	II. 74
5.5 Listing of Printer Output.....	II. 83

6. CASE STUDIES AND AN APPLICATION.....	II. 91
6.1 Introduction: Models I, II, III and IV.....	II. 91
6.2 Seismic Risk Spectra at 6 Sites.....	II. 95
6.3 Comparison of Different Models.....	II. 95
6.4 An Application.....	II.106
ACKNOWLEDGEMENTS.....	II.119
REFERENCES.....	II.120

1. INTRODUCTION

Earthquakes are natural phenomena occurring in many parts of the world and can adversely affect all societies. Even though the annual losses due to earthquakes are smaller than those caused by other natural disasters like wind and flood, it is the unforseeable nature and destructive power of a major earthquake that commands serious attention. With rapidly evolving modern technology the society's vulnerability to earthquakes will increase with time, as our activities become more dependent on new sources of power, communication and complex technological systems. Earthquakes can cause extensive loss of lives through collapses of buildings, dams, bridges and other man-made structures. They are but one hazard against which design provisions must be made. To this end considerable emphasis on earthquake prediction research in the U.S. has been initiated by seismologists in the 1970's. Up till now, the development of earthquake prediction techniques is far from complete. However, we note that even a successful prediction cannot eliminate the physical damages of an earthquake. Even if the entire population were to be evacuated to safety and, in time, the destruction of the structures that remain behind could still result in a major loss and a serious blow to the region's economy. It is thus necessary to apply effectively the knowledge of balanced earthquake resistant design through seismic risk analyses, to minimize the collapse or destruction of structures, through rational and economically balanced design codes.

Seismic risk analyses are capable of estimating the probability of experiencing a given earthquake of given size during the expected life of a structure. It should be noted that designing an earthquake resistant structure may add substantially to the cost of construction, and yet the

probability that any given structure will ever be adversely affected by a major earthquake is typically low. Earthquake resistant design thus aims at obtaining an optimum balance between the expected benefits and increased costs of construction.

The objective of a seismic risk analysis is to describe the nature and probability of possible future shaking. Methods for this evaluation have been developed and used as a useful engineering and seismological tool for over a decade now. As pointed out earlier, much work still remains to be done before seismologists can predict, accurately the large future earthquakes which will occur and cause strong shaking at a given site. The same applies for providing a precise description of what the shaking will be. The locations and characteristics of future earthquakes affecting a region are, in general, still in many ways a "random process" and their analysis can only be formulated on a probabilistic basis. In any case, the problem involved is to derive, from available seismological and geological information, the nature of the ground motion to be expected at the site during possible future strong-motion earthquakes. This information is then used as the input for quantifying the seismic loads required in the design of structures.

Early analytical methods to determine the seismic risk at a site were discussed by Cornell (1968). The results are often presented in terms of one ground motion parameter, such as peak acceleration, as a measure of some characteristic of shaking, and the return period is frequently calculated versus that parameter. Other ground motion parameters like magnitude, some peak amplitude of ground shaking and Modified Mercalli Intensity at the site have also been used (Milne and Davenport, 1969; Liu and Fagel, 1972; Dalal, 1973; DeCapua and Liu, 1974; Douglas and Ryall,

1975; Algermissen and Perkins, 1976). In such studies, the detailed spectral nature of ground motion is not considered. The probability that such a spectral amplitude will be exceeded during future earthquakes does depend on wave frequency (McGuire, 1974; Trifunac, 1977).

Implicit in most early analyses is the assumption that the total energy released during an earthquake is released from the focal point of the earthquake, and hence there are "point-source models." Der-Kiureghian and Ang (1977) proposed a risk model based on the assumption that an earthquake originates at the focus and propagates as an intermittent series of fault ruptures or slips in the ruptured zone. The peak intensity of ground shaking at a site is determined by the slip that is closest to the site. Der-Kiureghian and Ang studied the level of risk independently for maximum acceleration, velocity and displacement and discussed how a response spectrum derived from the three peak values changes shape with changes in the seismicity distribution and levels of risk.

Anderson and Trifunac (1977) generalized these methods to a functional of shaking, $S(\omega)$, which can represent any functional of strong ground motion at frequency ω (like Fourier amplitude, response spectral amplitude, peak response amplitude of any particular instrument, or duration of strong shaking). Their work also incorporated a more realistic model for describing the seismicity and proposed two independent methods to obtain uniform risk functionals: one assuming that the seismicity which is the input to the model is treated as the mean of a Poisson sequence, and the other one assuming it be taken literally.

The present effort is an extension of previous analyses on seismic risk and represents the second generation of the results based on the previous work of Anderson and Trifunac (1977). It incorporates into the

model the uncertainties involved in the characterization of seismicity. It uses the new frequency-dependent attenuation function (Trifunac and Lee, 1985a), and the new method for scaling Fourier and response spectral amplitudes (Trifunac and Lee, 1985b,c).

2. MODEL OF SEISMICITY

Following Anderson and Trifunac (1977), the seismicity here refers to the description of where earthquakes occur, the frequency of their occurrence and sizes (magnitudes or intensities), and the determination of the largest events expected to occur in any particular region or on any particular fault. The uncertainties involved in such description are also considered.

2.1 Types of Faults

The spatial distribution of earthquakes in any region can be divided into different source geometries according to their shape and pattern of seismicity. For this purpose, five types of sources are considered: point source, line source, areal source (diffused zone), dipping plane source, and volume (diffused) source.

(a) Point Source

The point source is used to describe concentrated seismicity, such as a geothermal area or a distant volcanic source, or in general when the source is far from the site.

(b) Line Source

A line source is used to model regions where the recorded seismic events lie fairly well along a line of shallow faults or when the faults are at considerable distance from the site.

(c) Areal Source (diffused zone)

An areal source describes areas where the faults are too numerous, randomly oriented and too small to describe individually, areas where the faults are not well defined, and where earthquakes have not been associated with faults.

These three types of sources all assume that the focal depth is small or unimportant for analysis and is thus taken as zero.

(d) Dipping Plane Source

A dipping source models the seismicity near large faults on which the activity is recorded over a depth range. The faults are modeled by planes which can be vertical or dipping at an angle.

It is assumed, in general, that the focus of an earthquake is equally likely to occur anywhere on a given fault surface.

(e) Volume Source

This model of the seismicity is to be employed where the relative depths of the earthquakes are important but cannot be constrained to a fault surface. It allows for the modeling of faults on which the activity is recorded over an area and within a certain depth. As pointed out by Anderson and Trifunac (1977), the general formulation of this type of source will become more important as more strong-motion data for deep earthquakes become available.

2.2 Estimation of Seismicity

There are several ways in which seismicity may be estimated. Which-ever method is used, the outcome of a seismic risk analysis will depend critically on such description of seismicity surrounding the site. In

a large enough region, the occurrence rate of earthquakes is often well known. But seismic risk analysis at a specific site is more sensitive to seismic activities near the site, particularly within 25 to 50 km (Trifunac and Anderson, 1977). For such small regions, "historical seismicity" based on felt reports is often incomplete and may not represent the true seismicity of the region. In such cases, the knowledge of fault slip rates or regional strain rates from plate tectonics theory may be used to estimate "geological seismicity." This helps to increase the reliability of the description of seismicity in a region (Anderson, 1979a).

For any single source zone, let $N(e)$ be the estimated number of earthquakes with size greater than or equal to e that is to be expected in a specified time period. Here e is a general scalar variable describing the size of the earthquake, such as magnitude, M , or maximum intensity $I_{\max}$, for example. It can also be used as a vector variable when more parameters are involved, like for example, magnitude, rupture length, seismic moment and stress drop.

For the case of scaling in terms of magnitude or maximum intensity, the frequency of occurrence of different magnitudes (Richter, 1958) versus magnitude, giving the probability of different earthquake sizes may be used. This states that in a given period of time, the occurrence of an earthquake in a source zone can be approximated by

$$\log_{10} N(M) = a - bM \quad (2.2.1)$$

where $N(M)$ is defined above and M could mean either magnitude or maximum intensity, "a" and "b" are parameters to be estimated, often from past seismicity. The earthquake size can be specified to be within a range $[M_{\min}, M_{\max}]$. More precisely, (2.2.1) can be rewritten as

$$\log_{10} \mathcal{N}(M) = \begin{cases} a - bM & M_{\min} \leq M \leq M_{\max} \\ 0 & M_{\max} < M \end{cases} \quad (2.2.2)$$

Here $M_{\min}$ denotes the smallest magnitude of concern to engineering (like 3 or 3.5), and $M_{\max}$ the upper bound magnitude which depends on each source region.

The parameters "a" and "b" can be estimated by regression analysis of the number and size of past earthquakes from the source zone in question. Using the relationship between the seismic moment rate and relative slip rate as given by Brune (1970), the parameter "a" can also be estimated indirectly from the seismic moment rate (Anderson, 1979a) in the area.

For the purposes of this discussion the estimation of seismicity thus involves the estimation of the parameters "a," "b" and " $M_{\max}$." It requires the use of historical and geological data combined with subjective judgment. Uncertainties in this estimation thus should be included in subsequent analysis. The parameters $M_{\max}$ and $\mathcal{N}(M)$ will be treated as random variables each described by an appropriate probability distribution function.

2.3 The Source Elements

For each source zone, starting with $\mathcal{N}(M)$, the estimated number of earthquakes with size greater than or equal to M , within the time period of (2.2.1) and a size M_j , we define $N(M_j)$ to be the estimated number of earthquakes with sizes inside $[M_j - \delta M_j/2, M_j + \delta M_j/2]$. Then

$$N(M_j) = \mathcal{N}(M_j - \delta M_j/2) - \mathcal{N}(M_j + \delta M_j/2) \quad (2.3.1)$$

As in Anderson and Trifunac (1977), the earthquake size M will be discretized. For scaling with magnitudes, one half of magnitude unit, will be used for basic interval. Earthquakes of size M_j will thus mean earthquakes of size within $[M_j-.25, M_j+.25]$. For scaling with intensities, the Modified Mercalli Intensities are already discretized. The discretized values ($M_j, j = 1$ to J) will be used in subsequent analysis.

The total seismicity of a region will be modeled by a superposition of different types of source zones in the region. To evaluate the risk, each of the zones will be divided into small source elements, and to each of these source elements, say the i -th element, we will assign n_{ij} , the estimated number of earthquakes of size M_j which occur in the i -th element within the chosen time period.

n_{ij} is derived from $N(M_j)$ for different geometries of source zones, as discussed in Section 2.1. The detailed manner in which the source parameters are related to the shaking at the site will affect the way n_{ij} is evaluated. One earlier approach is to consider only the epicentral distance from the fault to the site. In this case then, n_{ij} will represent the estimated number of earthquakes with epicenters in the i -th element. Such approach has been proposed and used by Cornell (1968), for example. Dalal (1973) and McGuire (1974) also used this form. For such approach, n_{ij} is defined as follows:

(a) Point Source: If there are $N(M_j)$ events of size M_j , and the point source is the i -th element, then $n_{ij} = N(M_j)$.

(b) Line Source: If the i -th element has a length L_i and is part of a line source of length L , then $n_{ij} = N(M_j)L_i/L$.

(c) Areal Source: If the i -th element has an area A_i , and is part of an areal source (diffused zone) of total area A , then $n_{ij} = N(M_j)A_i/A$.

(d) Dipping Surface: This case is analogous to that of an areal source. If the i -th element has area Σ_i , and is part of a dipping surface of total area Σ , then $n_{ij} = N(M_j)\Sigma_i/\Sigma$.

(e) Volume Source: If the i -th element has volume V_i , and is part of a volume source of total volume V , then $n_{ij} = N(M_j)V_i/V$.

Alternatively, if one considers not the epicentral distance but the distance to the closest point of rupture on the fault, for example, then n_{ij} represents the estimated number of earthquakes which have their closest point of rupture to the site in the i -th element. In this approach, both the magnitude and rupture length or area of the events are considered.

The assumptions that the rupture is unilateral and the direction of rupture may be random is often used with this approach. Der-Kiuregnian and Ang (1975), Douglas and Ryall (1975), and Anderson and Trifunac (1977) all used a similar approach. The following is a description of how n_{ij} is derived for each source type:

(a) Point Source: The formulation of the point source model is identical to that of the first approach.

(b) Line Source: The following formulation has been used by Anderson and Trifunac (1977) in their computer program "EQRISK." The line source of total length L is described by a set of straight lines, which are input as a consecutive sequence of coordinates in terms of their latitudes and longitudes. The program then represents the line source as a sequence of

N_L equally spaced points along the fault. The spacing $\delta L = L/(N_L - 1)$ between the points is small enough so that this is equivalent to discretizing the line source in terms of N_L points. For an earthquake of size M_j (magnitude of maximum intensity), it is assumed that the earthquake ruptures the fault along a length ℓ_j . Let $\ell_j = k_j \delta L$, so that the rupture zone takes up k_j segments of the discretized points. Then there are $N_L - k_j$ "allowable" points along the fault that have equal probability of being the midpoint of a rupture zone. Each of these "allowable" points will have $N(M_j)/(N_L - k_j)$ "allowable" events of size M_j . For each of these points, let the i -th discretized point (the i -th element) of the fault be the point on the rupture zone closest to the site where the risk is to be calculated. In this case the number $N(M_j)/(N_L - k_j)$ is attributed to the i -th element. The total number of events, n_{kj} , of the i -th element of size M_j is built up in this way. Events with rupture length ℓ_j larger than the fault length L will mean that only one of the discretized points on the fault, namely the point closest to the site, will have $N(M_j)$ as the expected number of events of size M_j .

(c) Areal Source: Since an areal source describes surface areas where the faults are too numerous or too small to describe individually (diffused zone), or where the faults are not well defined, the program "NEQRISK" (Anderson and Trifunac, 1977) simply divides the area into small source elements and assumes the earthquakes to have equal probability of occurrence in each source element, as in the first approach.

(d) Dipping Source: The formulation for a dipping plane source is similar to that for a line source. First, the fault surface is divided into N_Σ equal size square elements with areas $\delta \Sigma = \Sigma/N_\Sigma$, where Σ is the

total area of the dipping source. The spacing $\delta\Sigma$ is small enough so that this is equivalent to discretizing the dipping surface in terms of N_Σ points. For an earthquake of size M_j , it is assumed that the earthquake ruptures the fault along an area a_j . Then an event with area a_j must rupture $a_j/\delta\Sigma$ adjacent fault elements. Let $a_j = k_j\delta\Sigma$, so that the rupture zone takes up k_j segments of the discretized points. The ruptured k_j segments will then be approximately square for small events and rectangular with the width equal to the width of the fault surface for larger events. As in the case of a line source, there are $N_\Sigma - k_j$ "allowable" points (or discretized areas) along the fault which have equal probability of being the focus of a rupture zone. Each of these "allowable" points will have $N(M_j)/(N_\Sigma - k_j)$ "allowable" events of size M_j . For each of these points, let the i -th discretized point (i -th element) of the surface be the point on the rupture zone closest to the site. Then $N(M_j)/(N_\Sigma - k_j)$ events are attributed to that element. The total number of events, n_{ij} , of the i -th element of size M_j is built up in this way. Note that the number of "allowable" locations, $N_\Sigma - k_j$, depends on the size of the rupture area and hence on the size of the events.

(e) Volume Source: As mentioned in Anderson and Trifunac (1977), there were, and there still are at present, insufficient empirical results to formulate methods of modifying the distribution of epicenters to account for the finite dimensions of rupture in a source zone of this type. For such source zone at sufficient distance from the site, the same characterization of volume source as that of the first approach is used here.

3. ATTENUATION OF RISK FUNCTIONALS FROM THE SEISMIC SOURCES

3.1 Previous Work

An example of previous work on a description of the attenuation of risk functionals from seismic sources used in seismic risk analysis is found in Anderson and Trifunac (1977). The risk functionals that have been used include Fourier spectral amplitudes, $FS(T)$, pseudo relative velocity spectral amplitudes, $PSV(T)$, relative velocity spectral amplitudes, $SV(T)$ and absolute acceleration spectral amplitudes, $SA(T)$. For scaling of Fourier spectral amplitudes $FS(T)$, using earthquake magnitude, source to station distance and local geology (parameter s or h), either the model of Trifunac (1976) or Trifunac and Lee (1978) have been used. The first model is given by (Trifunac, 1976):

$$\log_{10} FS(T)_{,l} = \begin{cases} M + \log_{10} A_0(R) - a(T)p_l - b(T)M_{\min} - c(T) - d(T)s \\ \quad - e(T)v - f(T)M_{\min}^2 - g(T)R, & M \leq M_{\min} \\ M + \log_{10} A_0(R) - a(T)p_l - b(T)M - c(T) - d(T)s \\ \quad - e(T)v - f(T)M^2 - g(T)R, & M_{\min} \leq M \leq M_{\max} \\ M_{\max} + \log_{10} A_0(T) - a(T)p_l - b(T)M_{\max} - c(T) - d(T)s \\ \quad - e(T)v - f(T)M_{\max}^2 - g(T)R, & M_{\max} \leq M \end{cases} \quad (3.1.1)$$

where $T(=2\pi/\omega_n)$ is the period, M is the local magnitude of the earthquake, $\log_{10} A_0(R)$ is a description of the attenuation with epicentral distance R as used by Richter (1958) to define the local magnitude, and p_l is a linear approximation for $0.1 < p_l < 0.9$, to the actual probability, p_a , say, that the spectral amplitude, $FS(T)$, will not be exceeded. s is a site condition of the local geology, $s = 0$ for alluvial sites, 2 for sites on hard basement (igneous) rocks and 1 is used for sites which are on consolidated sediments or are difficult to classify because of confusing geological environment. v is a component variable set to 0 for horizontal

motion and to 1 for vertical motion. The values for $M_{\min}$ and $M_{\max}$ are given by

$$\begin{aligned} M_{\min} &= -b(T)/2f(T), \quad \text{and} \\ M_{\max} &= (1-b(T)/2f(T)) . \end{aligned} \quad (3.1.2)$$

The resulting spectral amplitudes grow linearly for $M < M_{\min}$, quadratically for $M_{\min} \leq M \leq M_{\max}$ and stop increasing for $M > M_{\max}$. The coefficients $a(T)$, $b(T)$, ..., $g(T)$ are empirical "constants" estimated by regression analysis.

A more refined correlation function has been developed by Trifunac and Lee (1978). For $M_{\min} < M < M_{\max}$, it is given by

$$\log_{10} FS(T) = M + \log_{10} A_0(R) - b(T)M - c(T) - d(T)h - e(T)v - f(T)M^2 - g(T)R . \quad (3.1.3)$$

Here h represents the depth of alluvium, and replaces parameter s . The term $a(T)p_\ell$ is dropped from this correlation, and the result is the mean estimated value of $FS(T)$, $\hat{FS}(T)$. With $FS(T)$ representing the actual Fourier amplitude spectra computed from recorded accelerograms, the residuals, $\epsilon(T)$, were calculated, where

$$\epsilon(T) = \log_{10} FS(T) - \log_{10} \hat{FS}(T) \quad (3.1.4)$$

describes the distribution of the observed $FS(T)$ about the estimated $\hat{FS}(T)$. It is assumed that $\epsilon(T)$ can be described by a normal distribution function with mean $\mu(T)$ and standard deviation $\sigma(T)$ (Trifunac and Lee, 1978):

$$p_a = p(\epsilon, T) = \frac{1}{\sigma(T)\sqrt{2\pi}} \int_{-\infty}^{\epsilon(T)} \exp \left[-\frac{1}{2} \left(\frac{x - \mu(T)}{\sigma(T)} \right)^2 \right] dx, \quad (3.1.5)$$

from which p_a , the probability that $\log_{10}[FS(T)] - \log_{10}[\hat{FS}(T)] \leq \epsilon(T)$, can be determined directly.

Equations (3.1.1) and (3.1.3) have complementary uses, since in some cases, the depth of alluvium, h , is known, while in other cases, only the surface geology parameter s is available.

The regression analyses for the above equations were carried out for 186 free-field records corresponding to a total of 558 components of data from 57 earthquakes from 1933 through 1971. The same data base has been used for regression analysis on relative velocity spectral amplitudes, $SV(T)$ (Trifunac and Anderson, 1978b), on pseudo relative velocity spectral amplitudes, $PSV(T)$ (Trifunac and Anderson, 1978a; Trifunac and Lee, 1979), and absolute acceleration spectra amplitudes, $SA(T)$ (Trifunac and Anderson, 1977).

The same scaling equations were used for $SV(T)$, $SA(T)$ and $PSV(T)$ as those of $FS(T)$. In each case, the residuals, $\epsilon(T)$, were calculated, and then described by an analytical assumed probability distribution function. For $PSV(T)$, for example, the distribution function derived from Rayleigh distribution is used (Trifunac and Lee, 1979):

$$p_a = p(\epsilon, T) = [1 - \exp(-\exp(\alpha(T)\epsilon(T) + \beta(T)))]^{N(T)} \quad (3.1.6)$$

from which p_a , the probability that $\log_{10}[PSV(T)] - \log_{10}[\hat{PSV}(T)] \leq \epsilon(T)$, the difference between the calculated and estimated spectral amplitudes being less than a specified residual value, can be determined directly.

Scaling of Fourier spectral amplitudes in terms of Modified Mercalli Intensity and local geology parameter s can be found in Trifunac (1979), or in Trifunac and Lee (1978) using the depth of alluvium at recording sites, h . Similarly, scaling of pseudo relative response spectral amplitudes in terms of Modified Mercalli Intensity and local geology are found in Trifunac and Anderson (1978a) or Trifunac and Lee (1979). All these scaling relationships have been used in seismic risk analyses.

3.2 The New Scaling Functions

Through the years new earthquake acceleration data have been recorded, digitized and added to the original database. The list of 57 earthquakes has now grown to 104, most of which occurred in the regions of northern and southern California. The original list of 186 free-field records has now grown to 438. With 3 components available for each record, this amounts to a total of 1314 acceleration components, of which there are 876 horizontal and 438 vertical components.

With this new database, Trifunac and Lee (1985a) have developed a new frequency dependent attenuation function, $\mathcal{A}_{tt}(\Delta, M, T)$, to replace the Richter's attenuation function, $\log_{10} A_0(R)$, used previously. It takes the form

$$\mathcal{A}_{tt}(\Delta, M, T) = \mathcal{A}_0(T) \log_{10} \Delta, \quad (3.2.1)$$

where $\mathcal{A}_0(T)$ is an empirically determined parabolic function of period T , and Δ is given by

$$\Delta = S \left(\ln \frac{R^2 + H^2 + S^2}{R^2 + H^2 + S_0^2} \right)^{-1/2}. \quad (3.2.2)$$

Δ can be thought of as a "representative distance" from the earthquake source of size S , at depth H and at distance R from the recording site. S_0 is the coherence radius of the source. The definition of Δ used has been proposed by Gusev (1983) in his descriptive statistical model of earthquake source radiation for the description of short-period strong ground motion. Detailed description and calculation for this new attenuation function are given in Trifunac and Lee (1985a).

With the new attenuation function defined, the regression equation of Fourier amplitudes now takes the form (Trifunac and Lee, 1985b):

$$\log_{10} FS(T) = \mathcal{A}(\Delta, M, T) +$$

$$+ \begin{cases} M + C_1(T)M_{\min} + C_2(T)s + C_3(T)v + C_4(T) + C_5(T)M_{\min}^2, & M \leq M_{\min} \\ M + C_1(T)M + C_2(T)s + C_3(T)v + C_4(T) + C_5(T)M^2, & M_{\min} \leq M \leq M_{\max} \\ M_{\max} + C_1(T)M_{\max} + C_2(T)s + C_3(T)v + C_4(T) + C_5(T)M_{\max}^2, & M_{\max} \leq M \end{cases}$$

$$(3.2.1)$$

with all the parameters defined as before. Again, $M_{\min}$ and $M_{\max}$ are given by

$$M_{\min} = -C_1(T)/2C_5(T), \quad \text{and}$$

$$M_{\max} = -(1+C_1(T))/2C_5(T) \quad (3.2.2)$$

$C_1(T), \dots, C_5(T)$ are "coefficients" determined by regression analyses on the new database of 1314 components of Fourier amplitude data $FS(T)$ at 91 discrete periods T ranging from 0.04 to 15.0 sec. The same regression analysis has been also carried out with the site condition s replaced by the depth of alluvium h . The residuals $\epsilon(T)$ were again calculated as in the previous

analyses (equation (3.1.4)) and it was assumed that $\epsilon(T)$ can be described by a normal distribution function with mean $\mu(T)$ and standard deviation $\sigma(T)$ (equation (3.1.5)).

The same scaling equations in terms of the frequency dependent attenuation functions $\text{Att}(\Delta, M, T)$ were also developed for PSV(T) (Trifunac and Lee (1985c)). As before, the residuals, $\epsilon(T)$, were calculated and described by the appropriate distribution function as in equation (3.1.6).

Scaling of Fourier spectral amplitudes in terms of the Modified Mercalli Intensity (MMI) using the new database of 1314 records has also been carried out. While new instrumentation is being developed in many parts of the world, the long-term historical seismicity records continue to be available only in terms of the locally developed intensity scales and thus all risk calculations in terms of MMI scaling have been developed here as well.

The new set of earthquakes that have been added to the database through the years 1972 to 1981 have magnitudes typically below 6. For this reason the MMI levels for many of these earthquakes are either not well documented or have not been reported. A correlation of the MMI levels with the corresponding earthquake magnitudes, representative source to station distance and local site geology has been developed (Lee and Trifunac, 1985), using the following equation

$$I_{MM} = 1.5M - A - B \ln \Delta - C\Delta/100 - Ds, \quad (3.2.3)$$

where the parameters M , Δ , s are defined as before, and I_{MM} is the MMI level at the site.

For scaling of Fourier amplitude spectra the following equation was employed

$$\log_{10} FS(T) = b_1(T)\hat{I}_{MM} + b_2(T)s + b_3(T)v + b_4(T) \quad (3.2.4)$$

with $\hat{I}_{MM}$ representing the estimated level at the site computed from (3.2.3) or the reported MMI level if available. The correlation using h , the depth of alluvium, instead of s , the site parameters, was also carried out.

Correlations of pseudo relative velocity spectra in terms of intensities were also carried out (Trifunac and Lee, 1985c). In both bases, correlations using both the recording site conditions ($s = 0, 1, 2$) and the depth of alluvium, h , at the site have been presented.

In summary, for both Fourier and pseudo relative velocity response spectra, correlation coefficients for the four models using the new database are available: (1) magnitude- s model, (2) magnitude- h model, (3) MMI- s model and (4) MMI- h model.

3.3 Comparison of the Old and New Scaling Functions

It is interesting to compare the results of the old scaling functions using Richter's attenuation function and the old database with those of the new scaling functions using the new attenuation function and the new database. Figure 3.3.1 is a plot of the estimated PSV spectra with 5% of critical damping for an event of magnitude 6.5 at a focal depth of 5 km. In this example the recording sites are assumed to have the depth of sediments of 2 km and to be at the epicentral distances of 5, 25, 50 and 100 km. The two graphs correspond respectively to the horizontal component of input (on the left) and vertical component (on the right). The solid lines on each graph represent the new results while the dashed lines correspond to old

ESTIMATED PSV SPECTRA, 5% DAMPING
 MAGNITUDE = 6.5 FOCAL DEPTH(KM) = 5.0 ALLUV. DEPTH(KM) = 2.0
 EPIC DIST (KM) = 5 25 50 100

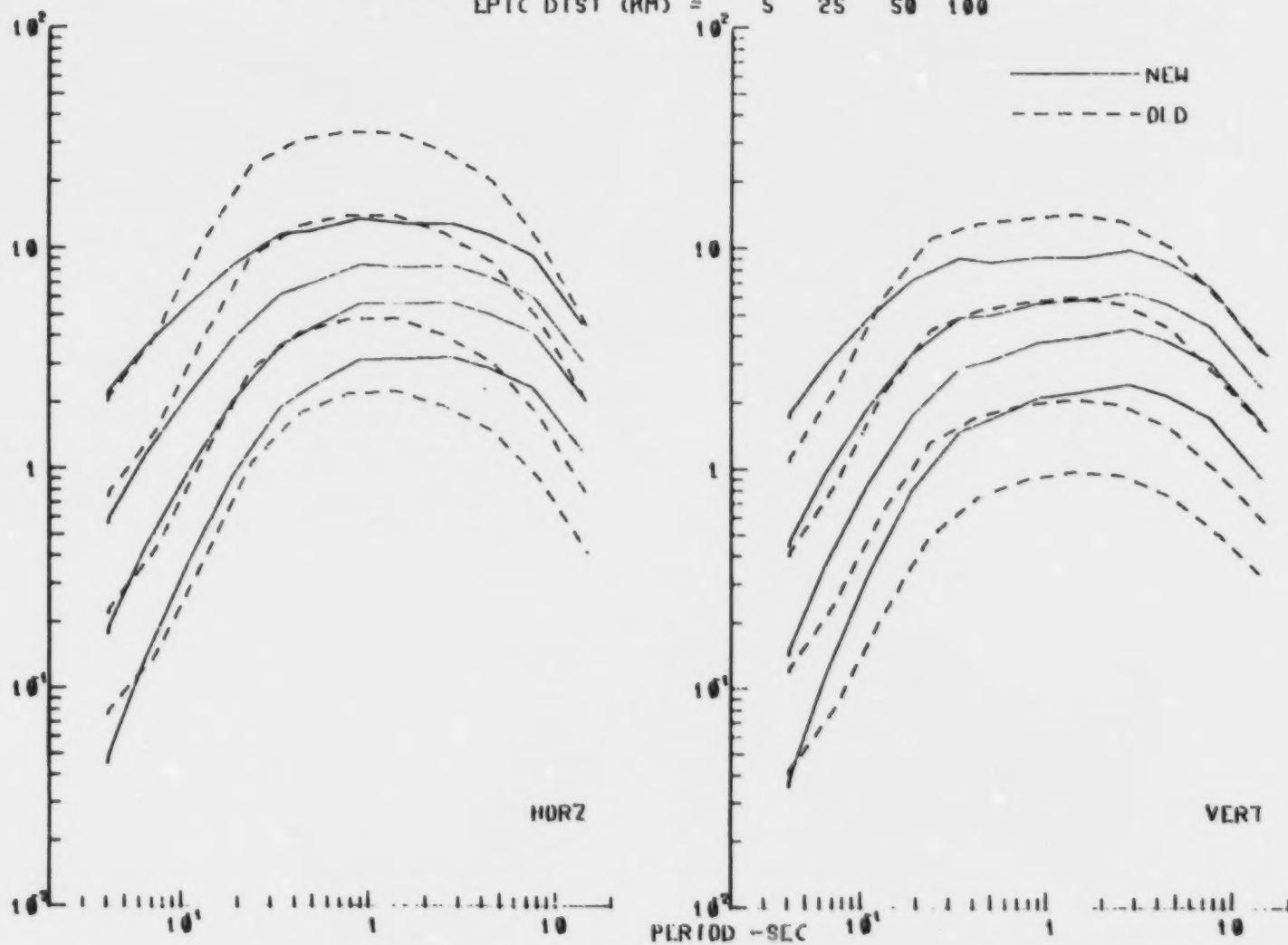


Figure 3.3.1

results. For periods between .04 and .1 seconds, the dashed lines and the corresponding solid lines are of comparable amplitudes. Beyond the period of .1 seconds, and for horizontal components the dashed lines are higher than the solid lines at sites with epicentral distances of 5 and 25 km, and vice versa at sites with epicentral distances of 50 and 100 km. For responses from the vertical accelerations at periods beyond 1 sec., the dashed line is higher than the solid line at the site with epicentral distance at 5 km, but the two sets of curves become of comparable amplitudes at a distance of about 25 km and then the trend is reversed for distances of 50 and 100 km. The observed differences between the dashed lines and solid lines are due to the differences between the old (Richter) attenuation and the new frequency dependent attenuation function.

Figure 3.3.2 is a plot of the estimated PSV spectra with 5% of critical damping at a recording site with epicentral distance of 5 km from the source. Again the recording site is assumed to have the depth of sediments equal to 2 km. The events shown have magnitudes of 5.5, 6.0, 6.5 and 7, and a focal depth of 5 km. Both the spectra from the horizontal and vertical acceleration show that for intermediate periods, the dashed lines are usually higher than the corresponding solid lines for most magnitudes. This shows that at short epicentral distances, the new frequency dependent attenuation function attenuates slower than the old Richter's attenuation function for periods greater than about .1 sec. Figure 3.3.3 is a plot of the estimated PSV spectra similar to that of Figure 3.3.2, but with the recording site this time at an epicentral distance of 100 km. Contrary to Figure 3.3.2, Figure 3.3.3 shows that for both the spectra from the horizontal and vertical acceleration, and for periods greater than about .1 sec., the dashed lines have lower

ESTIMATED PSV SPECTRA, 5% DAMPING
 FOCAL DEPTH(KM) = 5.0 ALT. DEPTH(KM) = 2.0 EPIC DIST(KM) = 5.0
 MAGNITUDE = 5.5, 6.0, 6.5, 7.0

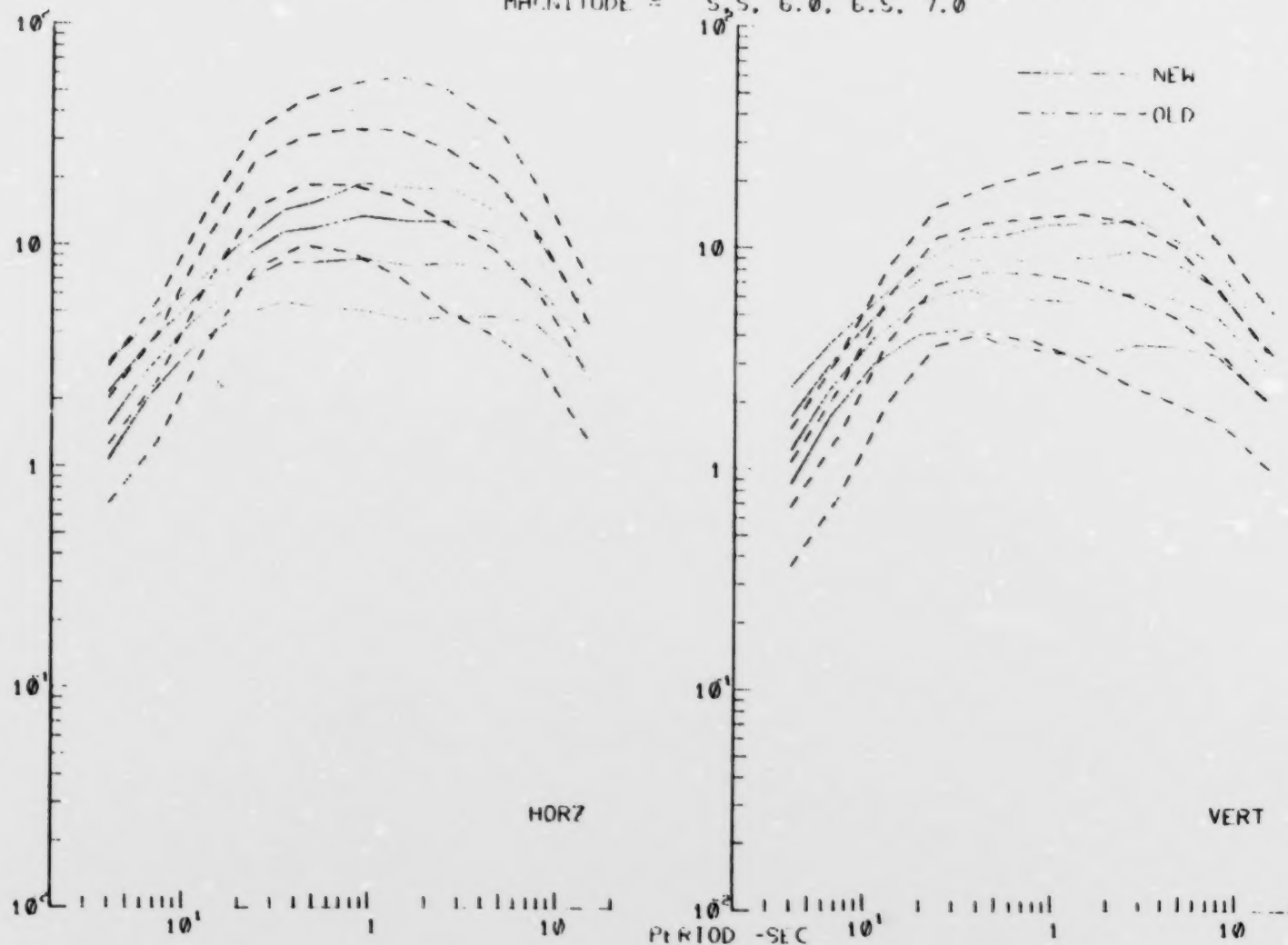


Figure 3.3.2

ESTIMATED PSV SPECTRA, 5% DAMPING
 FOCAL DEPTH(KM) = 5.0 ALLUV. DEPTH(KM) = 2.0 EPIC DIST(KM) = 100.0
 MAGNITUDE = 5.5, 6.0, 6.5, 7.0

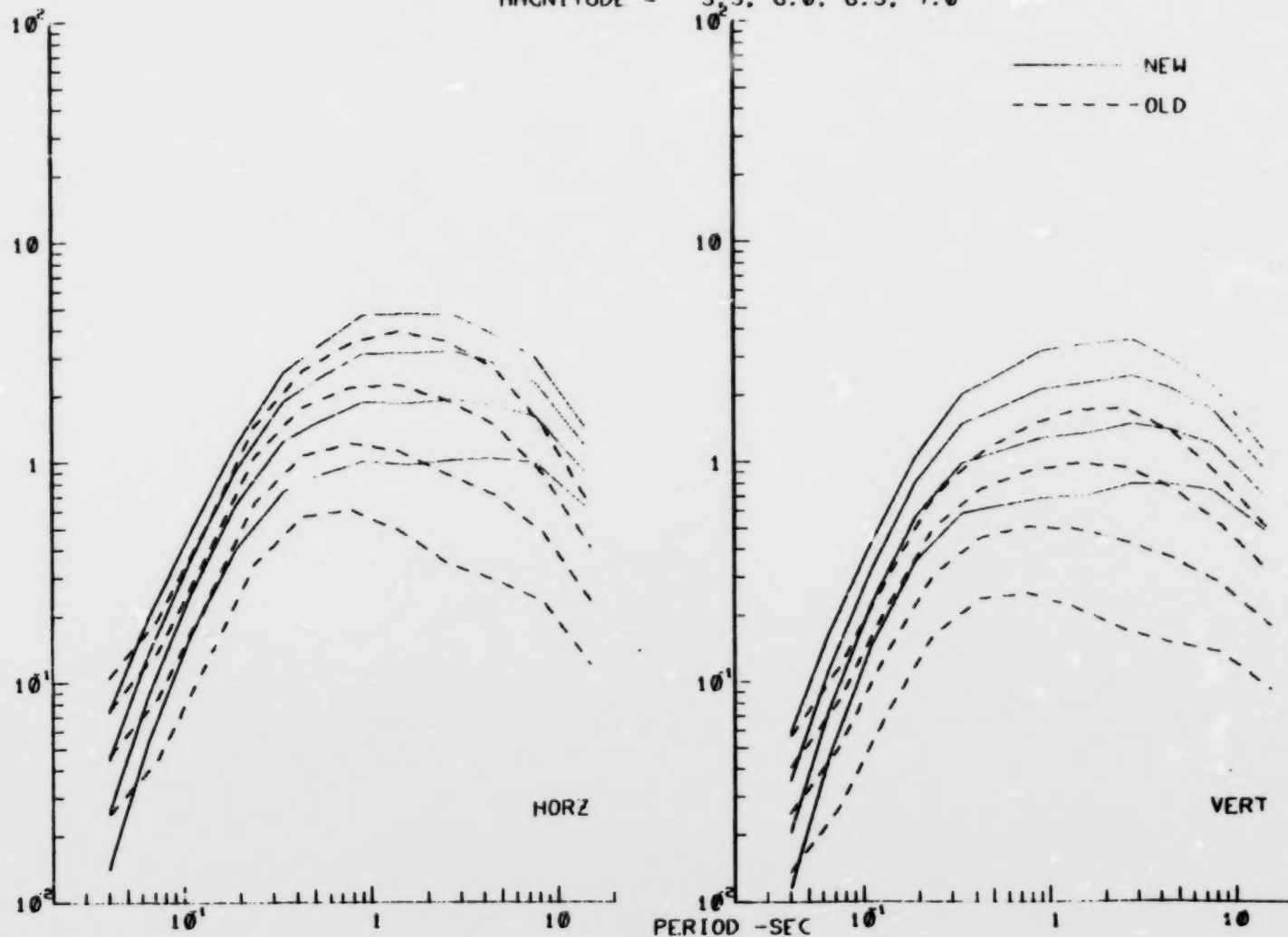


Figure 3.3.3

amplitudes than the corresponding solid lines for all magnitudes. At large epicentral distances, the new frequency dependent attenuation function attenuates more or less the same way as the old Richter's attenuation function.

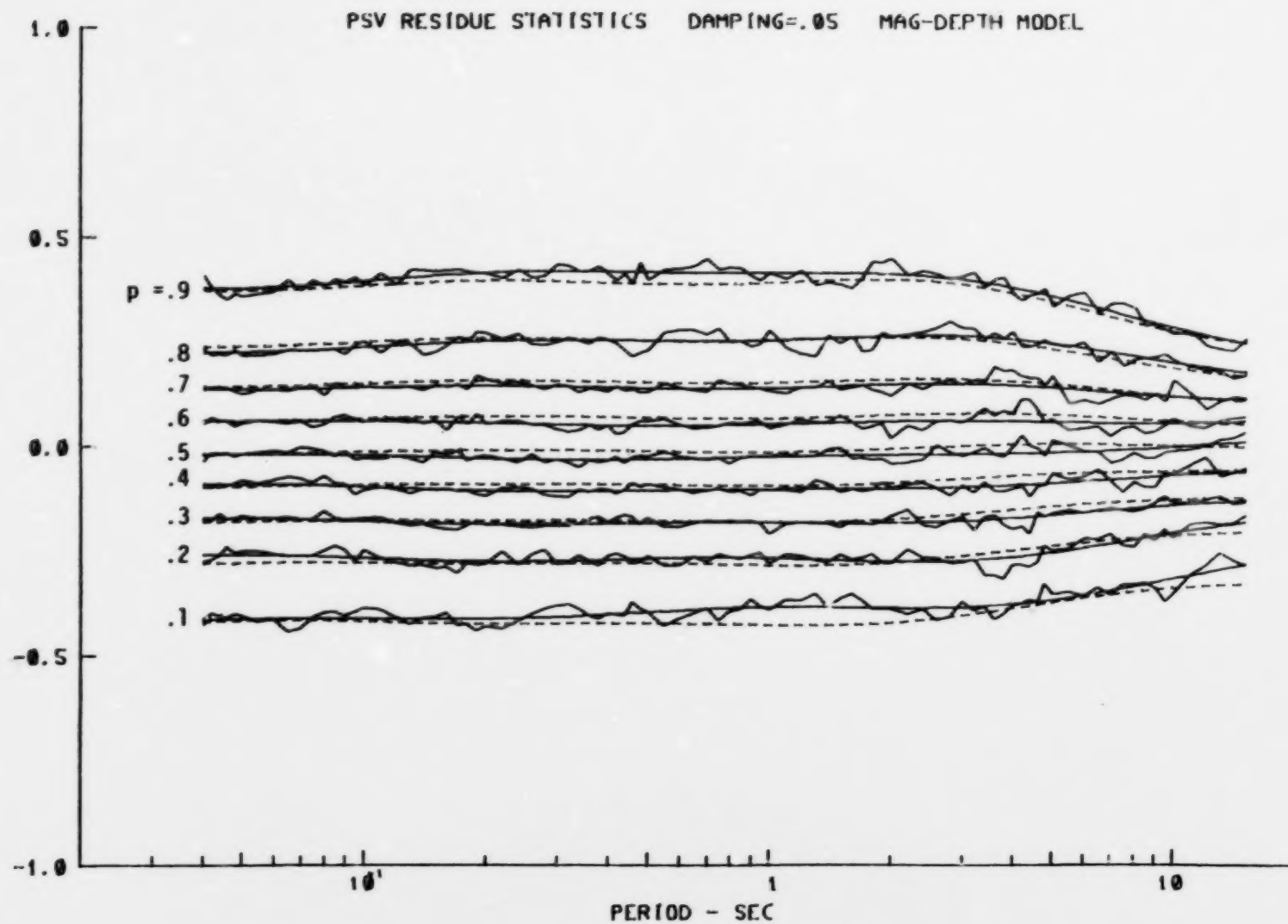
3.4 Derivation of the Function $q_{ij}(S(\omega))$

The empirical scaling functions summarized in the previous sections for Fourier and response amplitude spectra can now be applied and inverted to obtain the function $q_{ij}(S(\omega))$. For a given functional $S(\omega)$ of strong ground motion at frequency ω , $q_{ij}(S(\omega))$ is defined to be the probability that $S(\omega)$ will be exceeded at the site in an event of size M_j in the i -th source element. It implicitly assumes a scaling model of the attenuation of $S(\omega)$ from the source (the i -th element) to the site. Examples of these scaling models are presented in the previous sections. For a given model, let $\hat{S}_{ij}(\omega)$ be the estimated amplitude of the functional at the site from an event of size M_j in the i -th source element. Consider the residuals, ϵ_{ij} , defined by

$$\epsilon_{ij} = \log_{10} S(\omega) - \log_{10} \hat{S}_{ij}(\omega) \quad (3.4.1)$$

In the scaling model, it is assumed that ϵ_{ij} can be described by a probability distribution function, $p(\epsilon)$, from which p_{ij} , the probability that $\log_{10} S(\omega) - \log_{10} \hat{S}_{ij}(\omega) \leq \epsilon_{ij}$, can be determined directly. For Fourier amplitudes, for example, the normal distribution function is assumed, and for pseudo relative velocity spectral amplitudes, the distribution function is presented, as described in Section 3.1.

As an example, consider the scaling of PSV amplitudes at 5% of critical damping. Figure 3.4.1 shows the plot of the residual levels



11.24

Figure 3.4.1

corresponding to $p^*(\epsilon, T) = 0.1, 0.2, \dots, 0.8$ and 0.9 . The nine sets of curves, plotted versus period T , from the bottom to the top of the plot correspond to the residual levels at each of the probability levels, 0.1 through 0.9 . At each of the probability levels, the rough solid curve corresponds to the actual calculated residues at the particular level and the smooth solid curve is obtained by smoothing the rough solid curve along the T -axis. The corresponding dashed curve is the estimated residue, $\epsilon(T)$, at the particular level that is estimated by using the distribution function in (3.1.6). Both the calculated and estimated residue levels range from around $\epsilon = -0.4$ at $p^* = 0.1$ to around $\epsilon = +0.4$ at $p^* = 0.9$. The range of these residue levels is a measure of the distribution of the observed $PSV(T)$ about the estimated $\hat{PSV}(T)$.

Figure 3.4.2 shows a plot of the statistical parameters employed in the description of the residues from estimating PSV data of 5% damping. $N(T)$ and the smooth amplitudes of $\hat{\alpha}(T)$ and $\hat{\beta}(T)$ of (3.1.6) together with their 95% confidence intervals are respectively given in the top 3 plots of the figure. The two full curves in the bottom of the figure show the smoothed amplitudes of the computed χ^2 , $\chi^2(T)$ and Komolgorov-Smirnov, $KS(T)$, statistics, respectively. They are used to measure the goodness of fit. The dashed lines are their corresponding 95% cutoff levels. The reader is referred to Trifunac and Lee (1985b,c) for a more complete description of these tests. It is seen from the figure that both the χ^2 and K-S tests fail to reject the hypothesis that the distribution is that given in (3.1.6). The density function $p(\epsilon, T)$ thus represents an acceptable approximation to the actual $p^*(\epsilon, T)$.

Finally, the required function q_{ij} in the risk analysis is then defined by

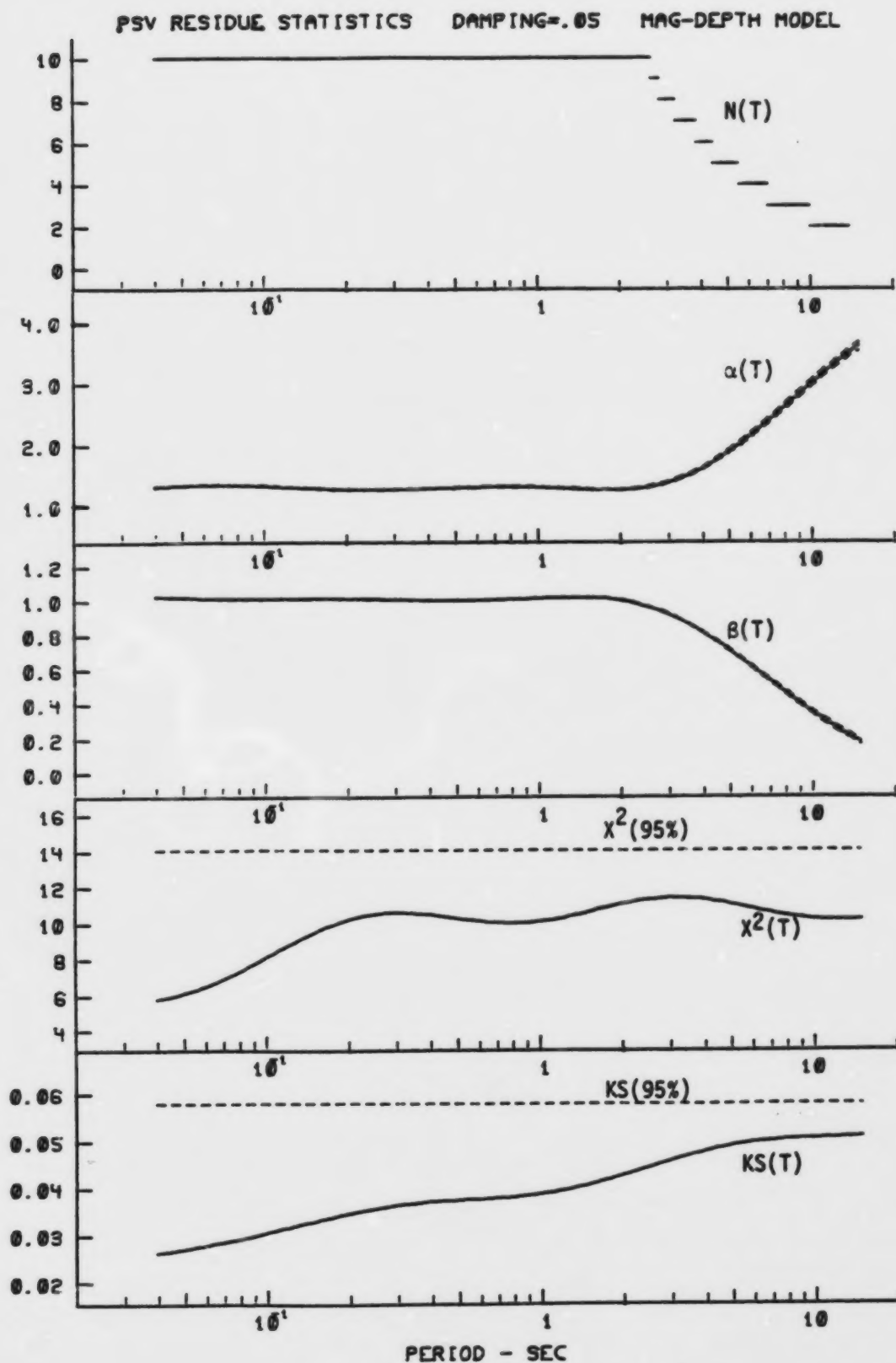


Figure 3.4.2

93

$$q_{ij} = 1 - p_{ij} \quad (3.4.2)$$

where p_{ij} is the estimated probability that $\log_{10} S(\omega) - \log_{10} \hat{S}(\omega) \leq \epsilon_{ij}$.

In summary, all the scaling relationships for Fourier and response spectral amplitudes that were derived by Trifunac (1976, 1979), Trifunac and Anderson (1978a,b), Trifunac and Lee (1978, 1979) and more recently by Trifunac and Lee (1985b,c) can also be used to find the function $q_{ij}(S(\omega))$ needed in risk calculations. The recent scaling relationships are an improvement over the previous ones by the same authors through the use of the new larger database and the use of a new frequency dependent attenuation function, as described in Trifunac and Lee (1985a).

As pointed out by Anderson and Trifunac (1977), the incorporation of the detailed description of scatter of amplitudes about the mean trend in the scaling equation, as is done, here, is one of the most significant improvements of our risk analysis over much of the previous work in this area. Der-Kiureghian (1977) pointed out that the uncertainty associated with the scaling functions often far exceeds the uncertainties associated with the other aspects of the modeling of seismic risk. Thus, by using the probability distribution functions as proposed here, a major source of potential errors in seismic risk analysis has been described and quantified.

4. MODEL OF SEISMIC RISK

4.1 The Uniform Risk Spectra

Consider a given functional $S(\omega)$ of strong ground motion and a given time interval of Y years at a given site. Following Anderson and Trifunac (1977), two functions of $S(\omega)$ and Y are to be determined:

- (1) The expected number of times, $N_E[S(\omega)]$, that $S(\omega)$ will be exceeded at the site in Y years, and
- (2) $p[S(\omega)]$, the probability that $S(\omega)$ will be exceeded at least once in Y years.

Figure 4.1.1 shows an example of the function $p[S(\omega)]$ at a particular frequency ω . If $p[S(\omega)]$ can be estimated at a chosen set of discrete frequencies, a uniform risk functional can be constructed as shown in Figures 4.1.2 and 4.1.3. Figure 4.1.2 shows the probability of exceeding $S(\omega)$ at three selected frequencies, ω_1 , ω_2 and ω_3 say. One can then construct a uniform risk spectra at a given probability level, $p = .1$, say, which at each frequency, ω , gives the spectral amplitude $S(\omega)$ that has the probability of 0.1 that it will be exceeded. To do this, one simply reads from these probability functions the spectral amplitudes S_i for each frequency ω_i which has probability 0.1 as in Figure 4.1.2 and plots the amplitudes S_i vs ω_i . Figure 4.1.3 is an example of such a plot, using only three frequencies for illustration purposes. In an actual calculation, one would use many frequencies and the resulting curve would have a much smoother appearance.

4.2 Formulation of the Risk Model

Let there be I source elements in the region of a zone. For the i -th source element, let n_{ij} be the estimated number of earthquakes of size M_j in the i -th element as described in Section 2. For the functional $S(\omega)$ at frequency ω , let q_{ij} be the probability that $S(\omega)$ will be exceeded at the site in an event of size M_j in the i -th element, as described in Section 3.

95

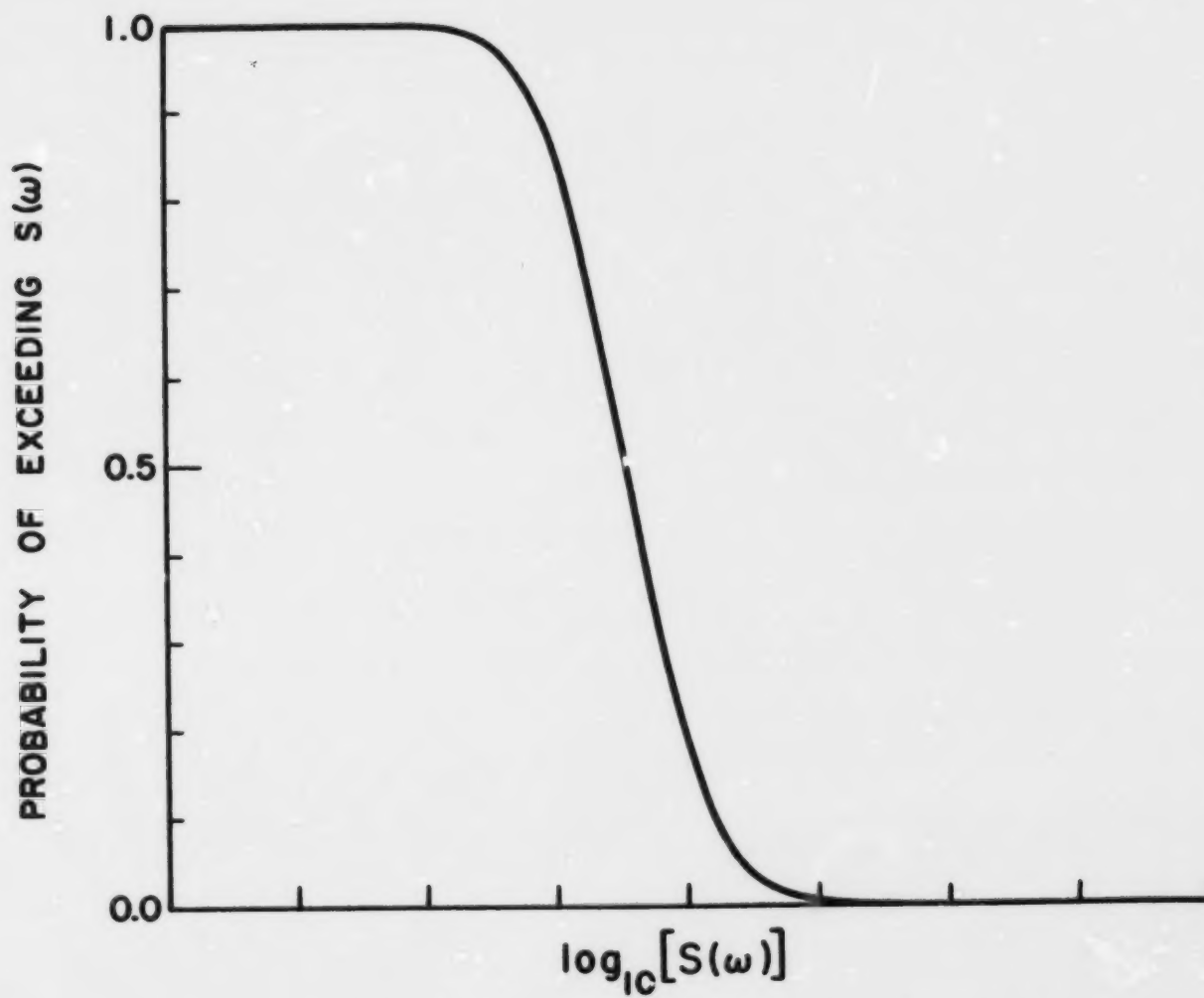


Figure 4.1.1 Example of the function $p[S(\omega)]$, the probability that $S(\omega)$ will be exceeded at least once in Y years.

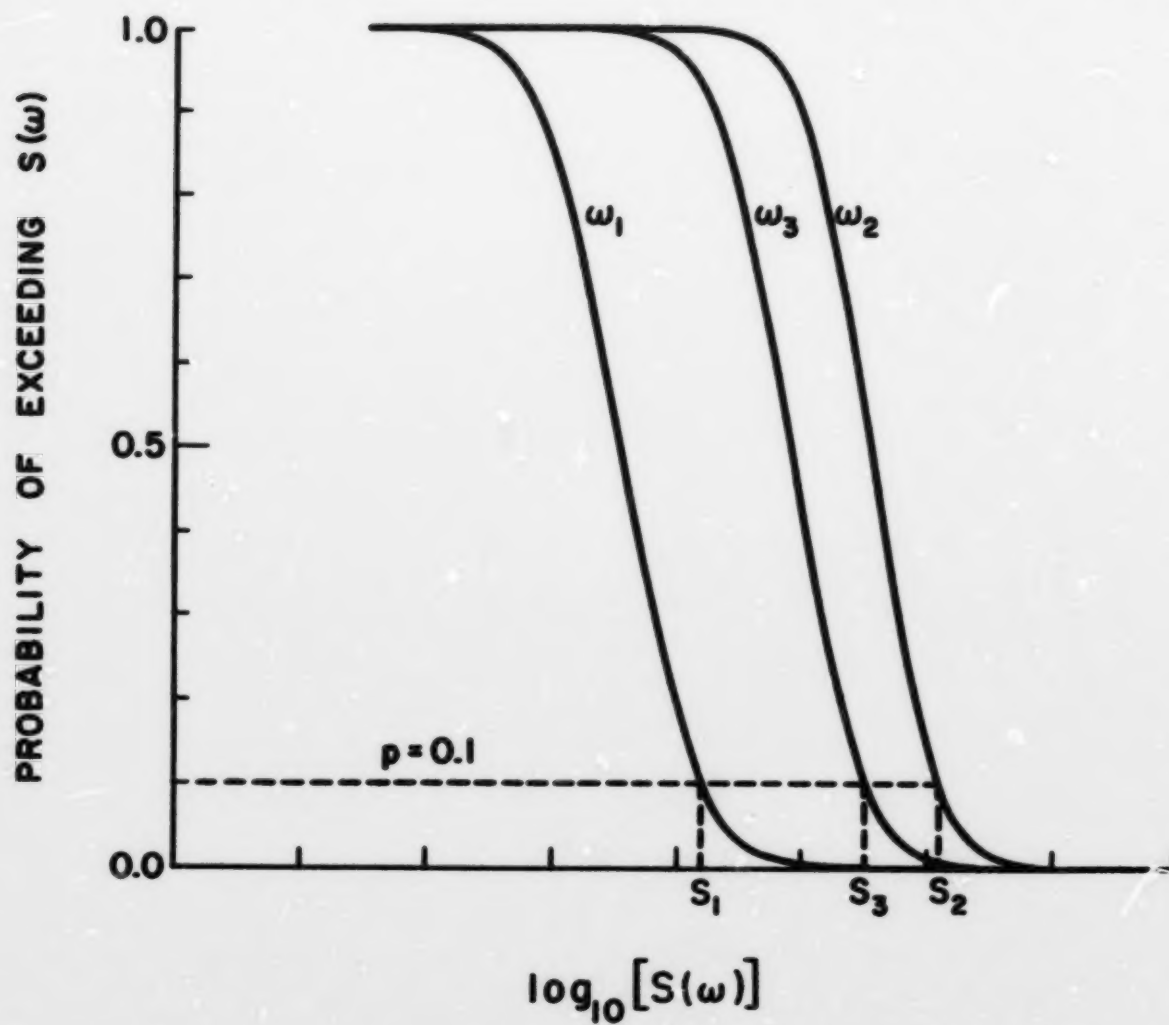


Figure 4.1.2 Example of the function $p[S(\omega)]$ at three different frequencies: ω_1 , ω_2 and ω_3 .

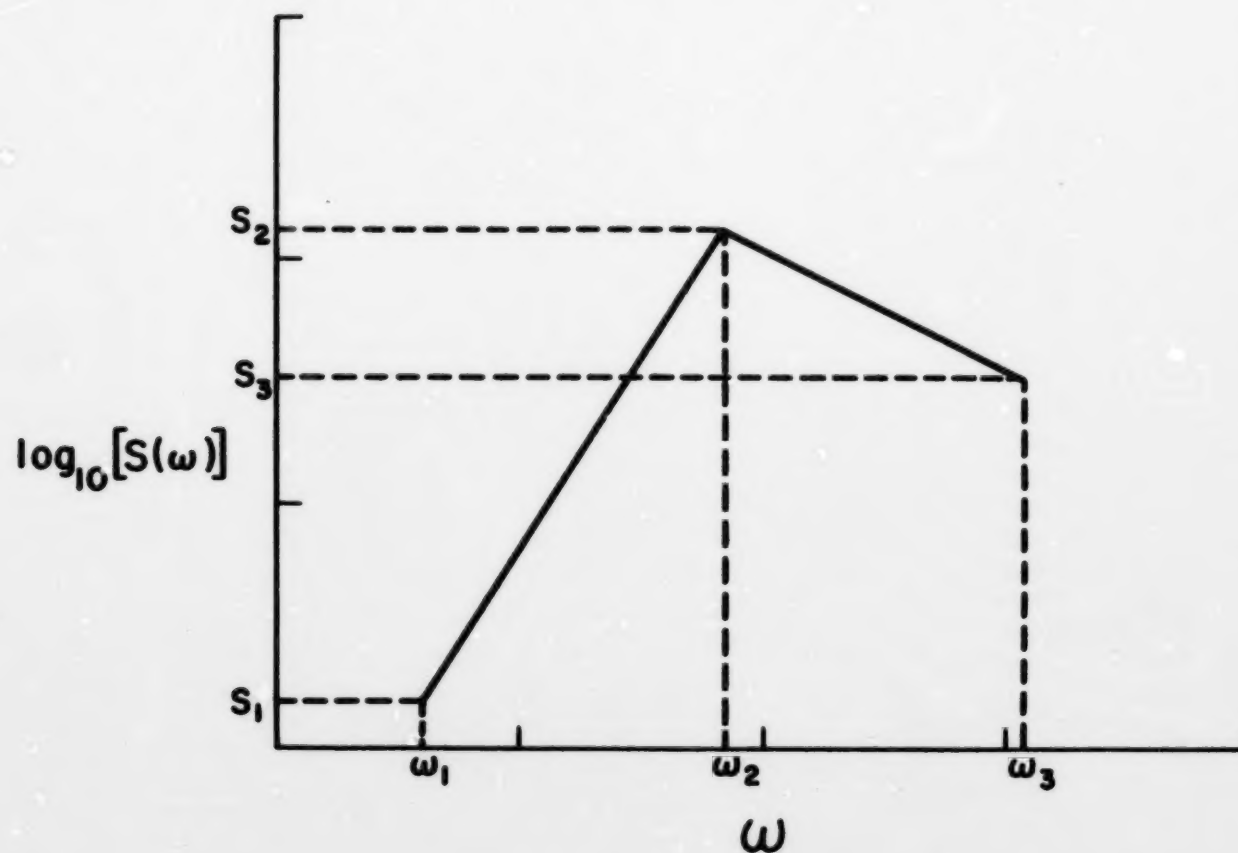


Figure 4.1.3 A uniform risk functional derived from the functions $p[S(\omega)]$ shown in Figure 4.1.2. At each frequency, the probability is 0.1 that $S(\omega)$ will be exceeded at least once in Y years. This figure uses only three frequencies for schematic purposes; in an actual calculation, one would use many frequencies and the resulting functional would have a smoother appearance.

Anderson and Trifunac (1977) derived $N_E(S(\omega))$ as follows. Assume, following Cornell (1968) and others, that n_{ij} is the mean of a Poissonian distribution. Then the probability of exactly k events of size M_j in the i -th element is $\exp(-n_{ij})n_{ij}^k/k!$. For k such events in the i -th element, the expected number of times that $S(\omega)$ will be exceeded at the site is kq_{ij} . Thus the expected number of times that $S(\omega)$ will be exceeded at the site from at least one event of size M_j in the i -th element, E_{ij} , is given by

$$E_{ij} = \sum_{k=0}^{\infty} kq_{ij} e^{-n_{ij}} n_{ij}^k/k! = q_{ij}n_{ij} . \quad (4.2.1)$$

Then, the expected number of times $S(\omega)$ will be exceeded at the site from all source elements, $i = 1$ to I , and all earthquake sizes M_j , $j = 1$ to J is given by

$$N_E[S(\omega)] = \sum_{i=1}^I \sum_{j=1}^J q_{ij}n_{ij} . \quad (4.2.2)$$

The above derivation assumes no uncertainty in the estimation of the seismicity of the i -th source element. As pointed out in Section 2, uncertainties in the estimation should be included in the risk analysis. Two parameters of the i -th source element, namely n_{ij} and $M_{\max}^i$, the maximum allowed (upper bound) earthquake size for events in the i -th element, will be treated as random variables each described by an appropriate probability density function.

As before, the conditional probability of exactly k events of size M_j in the i -th element, given n_{ij} , is $\exp(-n_{ij})n_{ij}^k/k!$. Since n_{ij} is a random variable described by a probability density function $p_n(n)$, say, the true unconditional probability of exactly k such events is given by

$$\text{prob}(k \text{ events of size } M_j) = \int_0^\infty \frac{e^{-n_{ij}} n_{ij}^k}{k!} p_n(n_{ij}) dn_{ij} \quad (4.2.3)$$

Again, the conditional probability that $S(\omega)$ will be exceeded at the site given an event of size M_j in the i -th source element is q_{ij} , given that $M_j \leq M_{\max}^i$. Thus the unconditional probability that $S(\omega)$ will thus be exceeded is $\overline{q_{ij}} = q_{ij} \cdot \text{Prob}(M_j \leq M_{\max}^i)$. For k such events, the expected number of times that $S(\omega)$ will be exceeded at the site becomes $k\overline{q_{ij}}$. As before, the expected number of times that $S(\omega)$ will be exceeded at the site from at least one event of size M_j in the i -th element, E_{ij} , is given by

$$\begin{aligned} E_{ij} &= \sum_{k=0}^{\infty} k\overline{q_{ij}} \int_0^\infty \frac{e^{-n_{ij}} n_{ij}^k}{k!} p_n(n_{ij}) dn_{ij} \\ &= \overline{q_{ij}} \int_0^\infty \sum_{k=0}^{\infty} \frac{kn_{ij}^k}{k!} e^{-n_{ij}} p_n(n_{ij}) dn_{ij} \\ &= \overline{q_{ij}} \int_0^\infty n_{ij} p_n(n_{ij}) dn_{ij} \\ &= \overline{q_{ij}} E(n_{ij}) \end{aligned} \quad (4.2.4)$$

as in (4.2.1). (4.2.4) is more general than (4.2.1) in that it includes the uncertainties in the estimation of seismicity at each chosen discrete earthquake size and the uncertainty in the estimated maximum allowed earthquake size. When the uncertainty in the estimation of seismicity is not present, $E(n_{ij})$ will be equal to n_{ij} . Similarly, when there is no uncertainty in the maximum allowed earthquake size, $\overline{q_{ij}}$ reduces to q_{ij} . Thus when both uncertainties are absent, (4.2.4) reduces to (4.2.1).

As before, the expected number of times that $S(\omega)$ will be exceeded at the site from all source elements and all allowed earthquake sizes is now given by

$$N_E[S(\omega)] = \sum_{i=1}^I \sum_{j=1}^J \overline{q_{ij}} E(n_{ij}) \quad , \quad (4.2.5)$$

where the summation i from 1 to I is over all source elements and the summation j from 1 to J is over all earthquake sizes.

Alternately, the same expressions for (4.2.4) and (4.2.5) can be derived by first assuming, as before, that the probability of exactly k events in the i -th element of size M_j is a Poissonian process. To include the uncertainties in the estimation of seismicity, the mean of the Poissonian distribution, n_{ij} , which is a random variable given by a probability density function $p_n(n)$, as described in the beginning of this section, can simply be taken as the expected value of n_{ij} , $E(n_{ij})$. Then the probability of exactly k events of size M_j in the i -th element becomes $\exp(-E(n_{ij}))E(n_{ij})^k/k!$. Proceeding as before, for k such events in the i -th element, the expected number of times that $S(\omega)$ will be exceeded at the site, including the uncertainty in the maximum allowed earthquake size is $k\overline{q_{ij}}$. Thus the expected number of times that $S(\omega)$ will be exceeded at the site from at least one event of size M_j in the i -th element, E_{ij} , is now given by

$$E_{ij} = \sum_{k=0}^{\infty} k\overline{q_{ij}} e^{-E(n_{ij})} E(n_{ij})^k/k! = \overline{q_{ij}} E(n_{ij}) \quad , \quad (4.2.6)$$

which is identical to (4.2.4). The expected number of times that $S(\omega)$ will be exceeded at the site from all source elements and all allowed

earthquake sizes, $N_E[S(\omega)]$, now follows from (4.2.6) the same way and is given in (4.2.5).

4.3 Characterizing Uncertainties of Seismicity

The previous section includes uncertainties in the estimation of seismicity and maximum allowed earthquake sizes in the calculation of the expected number of times that the spectral amplitude $S(\omega)$ will be exceeded at the site. This section illustrates examples of how such uncertainties can be estimated and used in the description of seismicity. As stated in Section 4.2, it is assumed that n_{ij} , the estimated number of earthquakes of size M_j in the i -th element, and $M_{\max}^i$, the maximum allowed (upper bound) earthquake size for events in the i -th element, are both treated as random variables, each described by an appropriate probability density function.

For each earthquake size M_j , let n_{ij} be described by a probability density function $p_n(n)$. As shown in Section 2.3, n_{ij} is estimated from $N(M_j)$ of the whole source zone of which the i -th element is a subelement. $N(M_j)$ is the expected number of earthquakes of sizes in a specified interval around M_j in the source zone. It in turn is estimated from an empirical scaling formula of the form, for $N_j = N(M_j)$,

$$\log N_j = a - bM_j, \quad (4.3.1)$$

where a and b are statistical parameters often estimated by regression.

The uncertainties in the estimation of $N(M_j)$ and hence n_{ij} are associated with those of a and b . For a given earthquake size M_j , $\log N_j$ and hence $\log n_{ij}$ can first be described by a probability density function. The following are examples of such functions.

(a) Suppose that for a particular earthquake size M_j and at a particular zone, $y = \log N_j$ is uniformly distributed with mean μ in the range $[\mu-\delta, \mu+\delta]$ (Figure 4.3.1a). The probability density function of y , $p_y(y)$, is given by

$$p_y(y) = \begin{cases} \frac{1}{2\delta} & \mu-\delta \leq y \leq \mu+\delta \\ 0 & \text{otherwise} \end{cases} \quad (4.3.2)$$

The probability density function of N_j , $p_n(n)$, is related to $p_y(y)$ through the relation $N_j = 10^y$, and is given by

$$p_n(n) = \begin{cases} p_y(y)/\frac{dn}{dy} & n \geq 0 \\ 0 & \text{otherwise} \end{cases} \quad (4.3.3)$$

with $dn/dy = d(10^y)/dy = c 10^y$, where $c = \ln 10$.

The uncertainties in the estimation of N_j are then included in the analysis by calculating $E(N_j)$ the expected value of N_j , which is to be used in (4.2.5). For this case, $E(N_j)$ is given by

$$\begin{aligned} E(N_j) &= \int_0^\infty n p_n(n) dn = \int_0^\infty 10^y (p_y(y)/\frac{dn}{dy}) dn \\ &= \int_{-\infty}^\infty 10^y p_y(y) dy = \frac{1}{2\delta} \int_{\mu-\delta}^{\mu+\delta} 10^y dy \\ &= \frac{1}{2c\delta} (10^{\mu+\delta} - 10^{\mu-\delta}) , \end{aligned} \quad (4.3.4)$$

II.37

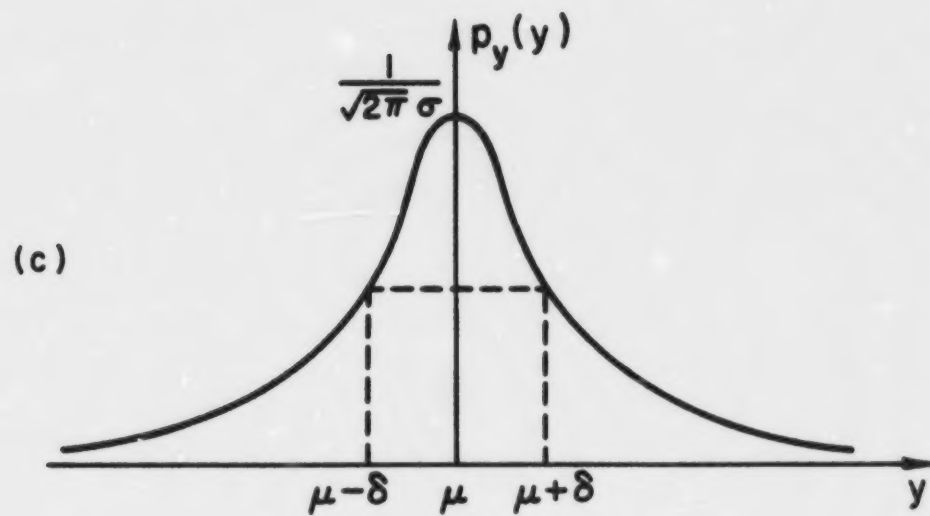
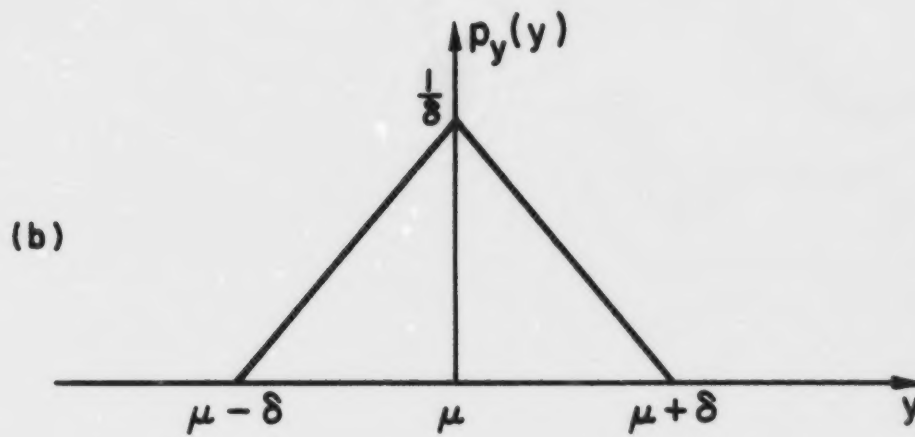
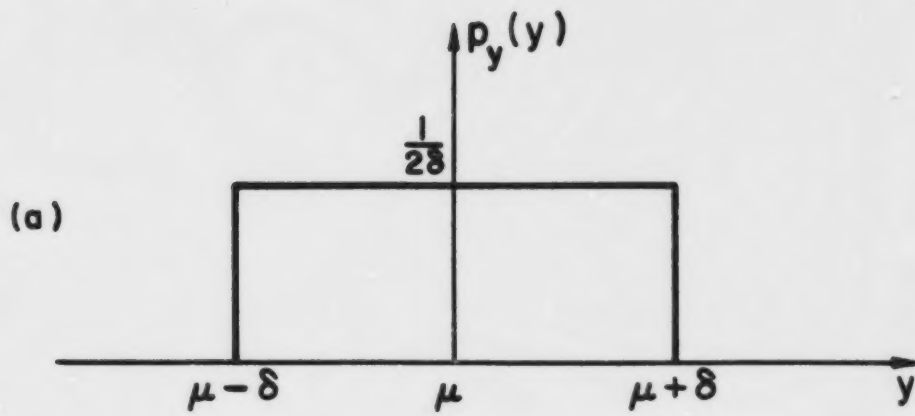


Figure 4.3.1

or

$$E(N_j) = 10^\mu \frac{\sinh c\delta}{c\delta}, \quad (4.3.5)$$

with $c = \ln 10$. Note that if $\delta = 0$, then $\mu = \log N_j$, and $E(N_j) = N_j = 10^\mu$.

(b) Suppose next that $y = \log N_j$ is distributed according to a triangular distribution with mean μ in the range $[\mu-\delta, \mu+\delta]$ (Figure 4.3.1b), so that the probability density function $p_y(y)$ is given by

$$p_y(y) = \begin{cases} (\delta+y-\mu)/\delta^2 & \mu-\delta \leq y \leq \mu \\ (\delta-(y-\mu))/\delta^2 & \mu \leq y \leq \mu+\delta \\ 0 & \text{otherwise} \end{cases} \quad (4.3.6)$$

The probability density function of N_j , $p_n(n)$, is then given by

$$p_n(n) = \begin{cases} p_y(y) \frac{dn}{dy} & n \geq 0 \\ 0 & \text{otherwise} \end{cases} \quad (4.3.7)$$

as in (4.3.3), with $dn/dy = c10^y$, $c = \ln 10$. $E(N_j)$ is now given by

$$\begin{aligned} E(N_j) &= \int_0^\infty np_n(n)dn = \int_{-\infty}^\infty 10^y p_y(y) dy \\ &= \int_{\mu-\delta}^\mu \frac{e^{cy}(\delta+(y-\mu))}{\delta^2} dy + \int_\mu^{\mu+\delta} \frac{e^{cy}(\delta-(y-\mu))}{\delta^2} dy \\ &= \int_{\mu-\delta}^{\mu+\delta} \frac{e^{cy}}{\delta} dy + \int_{\mu-\delta}^\mu \frac{e^{cy}(y-\mu)}{\delta^2} dy - \int_\mu^{\mu+\delta} \frac{e^{cy}(y-\mu)}{\delta^2} dy \\ &= \frac{e^{cy}}{c\delta} \Big|_{\mu-\delta}^{\mu+\delta} + \int_{-\delta}^0 \frac{e^{c\mu} e^{cx} x}{\delta^2} dx - \int_0^\delta \frac{e^{c\mu} e^{cx} x}{\delta^2} dx, \end{aligned} \quad (4.3.8)$$

or substituting $x = y - \mu$,

$$E(N_j) = \frac{e^{c\mu} 2 \sinh c\delta}{c\delta} - \frac{2e^{c\mu}}{\delta^2} \int_0^\delta x \cosh cx dx .$$

Integrating once again by parts,

$$E(N_j) = \frac{e^{c\mu} 2 \sinh c\delta}{c\delta} - \frac{2e^{c\mu}}{c\delta^2} (\delta \sinh c\delta - (\frac{\cosh c\delta - 1}{c})) . \quad (4.3.9)$$

The first two terms are identical and cancel. (4.3.9) becomes

$$\begin{aligned} E(N_j) &= \frac{2e^{c\mu} (\cosh c\delta - 1)}{c^2 \delta^2} \\ &= 10^\mu 2 \left(\frac{\cosh c\delta - 1}{c^2 \delta^2} \right) . \end{aligned} \quad (4.3.10)$$

Again $E(N_j) = 10^\mu$ if $\delta = 0$.

(c) As a third example, suppose $y = \log N_j$ is normally distributed with mean μ and standard deviation σ , so the distribution function of y , $p_y(y)$, is (Fig. 4.3.1c)

$$p_y(y) = \frac{1}{\sqrt{2\pi}\sigma} \exp[-(\frac{y-\mu}{\sigma})^2 / 2] . \quad (4.3.11)$$

$E(N_j)$ is now given by

$$\begin{aligned} E(N_j) &= \int_0^\infty np_n(n) dn = \int_{-\infty}^\infty 10^y p_y(y) dy \\ &= \int_{-\infty}^\infty \frac{1}{\sqrt{2\pi}\sigma} \exp[cy - \frac{1}{2} (\frac{y^2 - 2\mu y + \mu^2}{\sigma^2})] dy \end{aligned}$$

$$\begin{aligned}
&= \int_{-\infty}^{\infty} \frac{1}{\sqrt{2\pi}\sigma} \exp\left[-\frac{1}{2} \left(\frac{y^2 - 2(\mu + c\sigma^2) + (\mu + c\sigma^2)^2}{\sigma^2}\right)\right] \exp\left(\mu + \frac{c\sigma^2}{2}\right) dy \\
&= \exp\left(\mu + \frac{c\sigma^2}{2}\right) \int_{-\infty}^{\infty} \frac{1}{\sqrt{2\pi}\sigma} \exp\left[-\frac{1}{2} \left(\frac{y - (\mu + c\sigma^2)}{\sigma}\right)^2\right] dy
\end{aligned}$$

or
$$E(N_j) = 10^{\mu + c\sigma^2/2}, \quad (4.3.12)$$

with the last integral in (4.3.12) being equal to 1. Again $E(N_j) = 10^\mu$ when $\sigma = 0$.

In each of the above examples the uncertainties are characterized by two parameters, like the mean μ and the range $[\mu - \delta, \mu + \delta]$ in the first two examples of uniform and triangular distributions, or the mean μ and the standard deviation σ in the last example with normal distribution of $y = \log N_j$ at a particular earthquake magnitude or intensity. In each example, the mean of the distribution can be estimated, for example, by the regression equation as in (4.3.1), where for $N_j = N(M_j)$ instead of $\log N_j$, the mean $\mu = \mu(M_j)$ is estimated from the same empirical scaling formula

$$\mu(M_j) = a - bM_j, \quad (4.3.14)$$

with a and b the same statistical parameters as in (4.3.1). The uncertainties in the estimation of a and b can now be characterized by additional parameters δa and δb . They in turn are used to characterize the uncertainties of seismicity, $\delta = [\mu(M_j)]$, where

$$\delta[\mu(M_j)] = \delta[a - bM_j]. \quad (4.3.15)$$

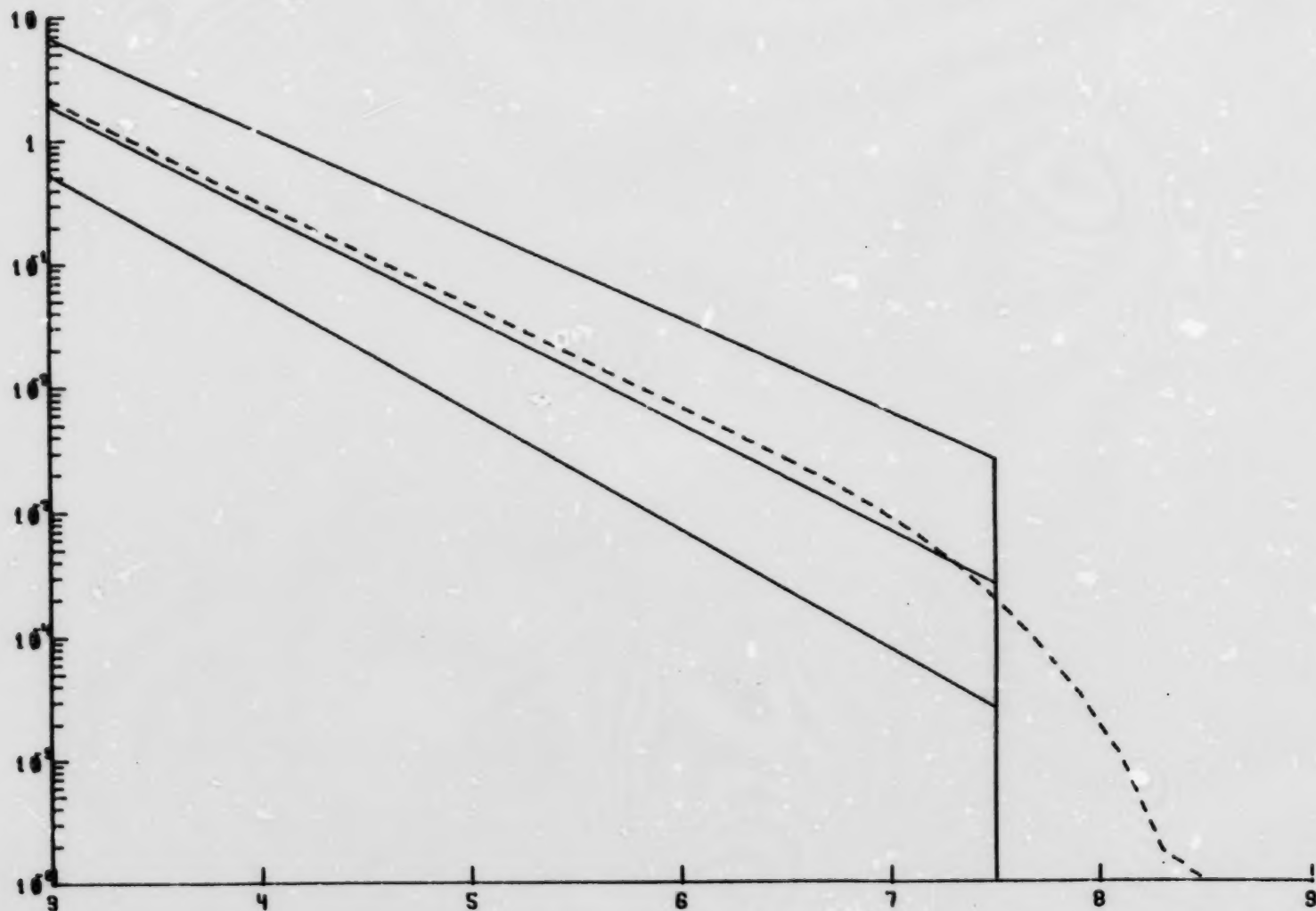
$\mu(M_j)$ and $\delta[\mu(M_j)]$ can now be used to estimate $E(N(M_j))$, the expected number of earthquakes of size M_j . Depending on the geometry of the zone, $E(n_{ij})$ of the i -th element of the zone can now be estimated from $E(N(M_j))$ as described in Section 2.3.

Figure 4.3.2 is an example of a plot of the seismicity at the Sierra Madre-Cucamonga fault in Los Angeles, California. It plots the number of earthquakes (on logarithmic scale) versus earthquake size in magnitude. The middle solid line represents the mean of the estimated number of earthquakes at each magnitude up to the maximum magnitude of 7.5. At each magnitude level, the estimated number of earthquakes is assumed to be triangularly distributed about the mean. The two outer solid lines represent the range of such distribution at each magnitude. The maximum magnitude $M_{\max}$ is also assumed to be a random variable having triangular distribution centered at 7.5 and in the range from 6.5 to 8.5. The dashed line in Figure 4.3.1 represents $E(N(M_j)) \cdot \text{prob}(M_j < M_{\max})$, a term used to include the uncertainties in the estimation of seismicity in the following seismic risk calculations.

4.4 The Probability of Exceedance $P[S(\omega)]$

In the previous sections we have discussed how the uncertainties in the estimation of seismicity and maximum allowed earthquake sizes could be included in the calculation of the expected number of times that the spectral amplitude $S(\omega)$ will be exceeded at the site. The function, $P[S(\omega)]$, the probability that $S(\omega)$ will be exceeded at least once in the given period, will next be derived. The following derivation is extended from that presented in Anderson and Trifunac (1977).

SIERRA MADRE - CUCAMONGA FAULT ZONE



11.42

Figure 4.3.2

Given the functional $S(\omega)$ at frequency ω , let:

p_{ij} = probability that at least one event of size M_j in the i -th element will cause $S(\omega)$ to be exceeded.

To find p_{ij} , the Poisson assumption is used, and again treating n_{ij} as a random variable described by a probability density function $p_n(n)$.

With $\bar{q}_{ij}$ defined as before,

$1 - \bar{q}_{ij}$ = probability of not exceeding $S(\omega)$ for one event,

$(1 - \bar{q}_{ij})^k$ = probability of not exceeding for k events,

$1 - (1 - \bar{q}_{ij})^k$ = probability of exceeding at least once for k events.

Combining this with (4.2.3), the probability of exceeding at least once for all k events of size M_j , gives

$$\begin{aligned} p_{ij} &= \sum_{k=0}^{\infty} (1 - (1 - \bar{q}_{ij})^k) \int_0^{\infty} \frac{e^{-n_{ij}} n_{ij}^k}{k!} p_n(n_{ij}) dn_{ij} \\ &= 1 - \int_0^{\infty} e^{-\bar{q}_{ij} n_{ij}} p_n(n_{ij}) dn_{ij} \\ &= 1 - E(e^{-\bar{q}_{ij} n_{ij}}), \end{aligned} \quad (4.4.1)$$

where $E(e^{-\bar{q}_{ij} n_{ij}})$ is the expected value of $e^{-\bar{q}_{ij} n_{ij}}$ under the probability distribution function $p_n(n_{ij})$. (4.4.1) is of the same form as given in Anderson and Trifunac (1977), where no uncertainties in seismicity and maximum earthquake size are assumed, so that $E(e^{-\bar{q}_{ij} n_{ij}}) = e^{-\bar{q}_{ij} n_{ij}}$. To evaluate the probability of exceedance, $P[S(\omega)]$, we proceed as follows:

$1 - p_{ij}$ = probability that no earthquake of size M_j at the i -th element will cause $S(\omega)$ to be exceeded,

$\prod_{j=1}^J (1-p_{ij})$ = probability that no earthquake of any size at the i -th element will cause $S(\omega)$ to be exceeded,

$1-P[S(\omega)] = \prod_{i=1}^I \prod_{j=1}^J (1-p_{ij})$ = probability that no earthquake anywhere in the source region will cause $S(\omega)$ to be exceeded.

Thus

$$\begin{aligned} P[S(\omega)] &= 1 - \prod_{i=1}^I \prod_{j=1}^J (1-p_{ij}) , \\ &= 1 - \prod_{i=1}^I \prod_{j=1}^J E(e^{-\bar{q}_{ij}n_{ij}}) , \end{aligned} \quad (4.4.2)$$

using (4.4.1). Assuming further that the earthquake events of all sizes in each element are independent, (4.4.2) simplifies to

$$\begin{aligned} P[S(\omega)] &= 1 - E\left(\prod_{i=1}^I \prod_{j=1}^J e^{-\bar{q}_{ij}n_{ij}}\right) \\ &= 1 - E\left(e^{\sum_{i=1}^I \sum_{j=1}^J \bar{q}_{ij}n_{ij}}\right) . \end{aligned} \quad (4.4.3)$$

(4.4.3) is in a form similar to (1.2) in Anderson and Trifunac (1977). When no uncertainty in seismicity is assumed, $E(e^{-\bar{q}_{ij}n_{ij}}) = e^{-\bar{q}_{ij}n_{ij}}$ and $\sum_{i=1}^I \sum_{j=1}^J \bar{q}_{ij}n_{ij} = N_E[S(\omega)]$, so that (4.4.3) reduces to

$$P[S(\omega)] = 1 - e^{-N_E[S(\omega)]} . \quad (4.4.4)$$

There is yet another way to derive $P[S(\omega)]$ under the assumption that the probability of having exactly k events in the i -th element of size M_j

is of Poisson type, and that both the uncertainties in the seismicity n_{ij} and maximum allowed earthquake size, $M_{\max}^i$, in the i -th element are also included. This is done by taking the mean of the Poisson distribution to be $E(n_{ij})$, the expected value of n_{ij} . The probability of exactly k events of size M_j occurring in the i -th element then becomes $\exp(-E(n_{ij}))E(n_{ij})^k/k!$ and p_{ij} takes the form:

$$p_{ij} = \sum_{k=0}^{\infty} (1 - (1 - \bar{q}_{ij})^k) \frac{\exp(-E(n_{ij}))E(n_{ij})^k}{k!}$$

$$= 1 - \exp(-\bar{q}_{ij}E(n_{ij})) \quad , \quad (4.4.5)$$

and (4.4.2) for $P[S(\omega)]$, the probability of exceedance, now takes the form

$$P[S(\omega)] = 1 - \prod_{i=1}^I \prod_{j=1}^J (1 - p_{ij}) \quad ,$$

$$= 1 - \exp\left(- \sum_{i=1}^I \sum_{j=1}^J \bar{q}_{ij} E(n_{ij})\right) \quad ,$$

$$= 1 - \exp(-N_E[S(\omega)]) \quad , \quad (4.4.6)$$

using (4.2.5) as the expression for $N_E[S(\omega)]$. As before, (4.4.6) shows the close relationship between the expected number of exceedances $N_E[S(\omega)]$, and the probability of exceedance, $P[S(\omega)]$, when a Poisson assumption is used to describe the seismicity rate, and here, even when the uncertainties in seismicity are included. (4.4.6) is a more convenient expression for calculating $P[S(\omega)]$ than (4.4.2) since it allows $P[S(\omega)]$ to be estimated from $N_E[S(\omega)]$ directly.

4.5 Literal Seismicity

The previous sections discussed the calculations of the expected number of times the given spectral amplitude $S(\omega)$ will be exceeded at the site and the probability that $S(\omega)$ will be exceeded at least once in a given period. It showed how the uncertainties in the estimation of seismicity and maximum allowed earthquake sizes can be included in the calculations. This section considers a different hypothesis when one assumes that an earthquake and its aftershock sequence of known magnitudes will occur someplace within a specified fault or a diffused region. In this case, one again needs to know the risk at various sites. The two functions $N_E^*[S(\omega)]$ and $P^*[S(\omega)]$ will now be derived with the assumption that one knows how many earthquakes will occur within the source zone. This is referred to as a "literal" model because the seismicity is to be interpreted literally. The following derivations are those given in Anderson and Trifuanc (1977) and are included here for completeness only.

Given a source region, suppose it is known that $N(M_j)$ events of size M_j for $j = 1$ to J will occur somewhere in the region, but where exactly in the region is not known. As in Section 2.3, the source region is now divided into I subelements. Let γ_{ij} be the probability that an event of size M_j occurs in the i -th element. With n_{ij} the expected number of events of size M_j in the i -th element as defined in Section 2.3, it is clear that

$$n_{ij} = \gamma_{ij} N(M_j) ,$$

$$\text{and} \quad \sum_{i=1}^I \gamma_{ij} = 1 , \quad (4.5.1)$$

$$\sum_{i=1}^I n_{ij} = N(M_j) = N_j .$$

There are many different ways in which the $N(M_j)$ events can be distributed among the I elements. To determine $N_E^*[S(\omega)]$, we proceed as in Section 4.2 to determine E_{ij}^* , the expected number of times that $S(\omega)$ will be exceeded at the site from at least one event of size M_j in the i -th element. E_{ij}^* is given by

$$E_{ij}^* = \sum_{n_{ij}=0}^{N_j} q_{ij} n_{ij} \cdot \text{prob} \left\{ \begin{array}{l} n_{ij} \text{ events of size } M_j \\ \text{occur in element } i \end{array} \right\}. \quad (4.3.2)$$

Considering the events as Bernoulli Trials, this is given by

$$\begin{aligned} E_{ij}^* &= \sum_{n_{ij}=0}^{N_j} q_{ij} n_{ij} \frac{N_j!}{n_{ij}!(N_j - n_{ij})!} \gamma_{ij}^{n_{ij}} (1 - \gamma_{ij})^{N_j - n_{ij}}, \\ &= q_{ij} N_j \gamma_{ij} = q_{ij} n_{ij}. \end{aligned} \quad (4.5.3)$$

Thus

$$N_E^*[S(\omega)] = \sum_{j=1}^J \sum_{i=1}^I E_{ij}^* = \sum_{j=1}^J \sum_{i=1}^I q_{ij} n_{ij}. \quad (4.5.4)$$

It is clear that $N_E^*[S(\omega)]$ is of the same form as $N_E[S(\omega)]$ and that if no uncertainties of estimation of seismicity are included in $N_E[S(\omega)]$, $N_E^*[S(\omega)]$ is identical to $N_E[S(\omega)]$.

To find $p^*[S(\omega)]$, consider:

$(1 - q_{ij})$ = probability that $S(\omega)$ is not exceeded from one event of size M_j in the i -th element,

$(1 - q_{ij})^{n_{ij}}$ = probability that $S(\omega)$ is not exceeded from n_{ij} events of size M_j in the i -th element,

$\prod_{i=1}^I (1-q_{ij})^{n_{ij}}$ = probability that $S(\omega)$ is not exceeded for a particular combination of events $\{n_{ij} | i \in I\}$ of size N_j .

The probability of any one particular combination of events is given, by the multinomial theorem, as

$$\frac{N_j!}{n_{1j}! n_{2j}! \dots n_{Ij}!} \gamma_{1j}^{n_{1j}} \gamma_{2j}^{n_{2j}} \dots \gamma_{Ij}^{n_{Ij}}, \quad (4.5.5)$$

and the sum of the probabilities of all combinations is, by the multinomial theorem, unity,

$$\begin{aligned} \sum \dots \sum \frac{N_j!}{n_{1j}! n_{2j}! \dots n_{Ij}!} \gamma_{1j}^{n_{1j}} \gamma_{2j}^{n_{2j}} \dots \gamma_{Ij}^{n_{Ij}} \\ = (\gamma_{1j} + \gamma_{2j} + \dots + \gamma_{Ij})^{N_j} = 1, \end{aligned} \quad (4.5.6)$$

where the summation is over all combinations of $\{n_{ij}, i=1 \text{ to } I\}$ such that $\sum_{i=1}^I n_{ij} = N_j$. Now the different combinations are mutually exclusive and exactly only one of them will occur, so the probability that $S(\omega)$ is not exceeded by any one combination is

$$\begin{aligned} \sum \dots \sum \left\{ \begin{array}{l} \text{probability that } S(\omega) \text{ is not} \\ \text{exceeded for a particular} \\ \text{combination} \end{array} \right\} \left\{ \begin{array}{l} \text{probability of that} \\ \text{particular combination} \end{array} \right\} \\ = \sum \dots \sum \left\{ \prod_{i=1}^I (1-q_{ij})^{n_{ij}} \right\} \left\{ \frac{N_j!}{n_{1j}! n_{2j}! \dots n_{Ij}!} \gamma_{1j}^{n_{1j}} \gamma_{2j}^{n_{2j}} \dots \gamma_{Ij}^{n_{Ij}} \right\} \\ = \sum \dots \sum \frac{N_j!}{n_{1j}! n_{2j}! \dots n_{Ij}!} [(1-q_{1j})\gamma_{1j}]^{n_{1j}} [(1-q_{2j})\gamma_{2j}]^{n_{2j}} \dots [(1-q_{Ij})\gamma_{Ij}]^{n_{Ij}} \end{aligned}$$

$$\begin{aligned}
&= \left\{ \sum_{i=1}^I [1-q_{ij}] \gamma_{ij} \right\}^{N_j} = \left\{ \sum_{i=1}^I \gamma_{ij} - \sum_{i=1}^I \gamma_{ij} q_{ij} \right\}^{N_j} \\
&= \left(1 - \sum_{i=1}^I \gamma_{ij} q_{ij} \right)^{N_j}, \quad (4.5.7)
\end{aligned}$$

where the summation in (4.5.7) is again over all combinations of $\{n_{ij} | i=1 \text{ to } I\}$ such that $\sum_{i=1}^I n_{ij} = N_j$. Since the different earthquake sizes M_j , $j=1$ to J are assumed to be independent,

$$\begin{aligned}
&\text{prob}\{S(\omega) \text{ is not exceeded by events of any size}\} = \\
&= \prod_{j=1}^J \text{prob}\{S(\omega) \text{ is not exceeded by events of size } M_j\} \\
&= 1 - p^*[S(\omega)], \quad (4.5.8)
\end{aligned}$$

so

$$p^*[S(\omega)] = 1 - \prod_{j=1}^J \left(1 - \sum_{i=1}^I \gamma_{ij} q_{ij} \right)^{N_j},$$

$$\text{or} \quad p^*[S(\omega)] = 1 - \exp \left\{ \sum_{j=1}^J N(M_j) \ln \left(1 - \sum_{i=1}^I \gamma_{ij} q_{ij} \right) \right\}. \quad (4.5.9)$$

Anderson and Trifuanc (1977) also derived an alternate set of expressions for $N_E[S(\omega)]$ and $p[S(\omega)]$, denoted by $N_E^+[S(\omega)]$ and $P^+[S(\omega)]$ under the same assumption that the number of future earthquakes $N(M_j)$ is to be taken literally. This is done by considering a very long time interval, say K times the base unit of Y years, or a duration of KY years in which $KN(M_j)$ events of size M_j will occur in the source regions. Further, in

the i -th element, it is assumed that $\gamma_{ij}KN(M_j) = Kn_{ij}$ events will occur, where Kn_{ij} is now an integer. For a sufficiently large number of events, this assumption will not introduce any significant error and is justifiable. Let

$N_E^{\#}[S(\omega)]$ = expected number of times that $S(\omega)$ will be exceeded during KY years,

$P^{\#}[S(\omega)]$ = probability that $S(\omega)$ will be exceeded at the site at least once in KY years.

These functions will first be derived from which $N_E^+[S(\omega)]$ and $P^+[S(\omega)]$ for Y years can be derived. Following the same logic, as in the derivation of $N_E[S(\omega)]$ and $N_E^*[S(\omega)]$,

$$N_E^{\#}[S(\omega)] = M \sum_{i=1}^I \sum_{j=1}^J n_{ij} q_{ij} \quad (4.5.10)$$

To derive $P^{\#}[S(\omega)]$, consider

$(1-q_{ij})$ = probability that $S(\omega)$ is not exceeded from 1 event of size M_j in the i -th element,

since Kn_{ij} is the number of events of size M_j that will occur in the i -th element in the duration of KY years,

$(1-q_{ij})^{Kn_{ij}}$ = probability that $S(\omega)$ is not exceeded from Kn_{ij} events of size M_j in the i -th element in KY years,

$\prod_{j=1}^J (1-q_{ij})^{Kn_{ij}}$ = probability that no earthquake of any size at the i -th element in KY years will cause $S(\omega)$ to be exceeded,

$$1 - P^{\#}[S(\omega)] = \prod_{i=1}^I \prod_{j=1}^J (1 - q_{ij})^{Kn_{ij}} = \text{probability that no earthquake any-}$$

where in the source region will cause $S(\omega)$ to be exceeded in KY years.

Thus

$$\begin{aligned} P^{\#}[S(\omega)] &= 1 - \prod_{i=1}^I \prod_{j=1}^J (1 - q_{ij})^{Kn_{ij}} \\ &= 1 - \exp\left\{ \sum_{i=1}^I \sum_{j=1}^J Kn_{ij} \ln(1 - q_{ij}) \right\}. \end{aligned} \quad (4.5.11)$$

Assuming that the K time intervals of Y years are independent, (4.5.10) and (4.5.11) give

$$N_E^+[S(\omega)] = \frac{1}{K} N_E^{\#}[S(\omega)] = \sum_{i=1}^I \sum_{j=1}^J n_{ij} q_{ij} = N_E^*[S(\omega)], \quad (4.5.12)$$

and

$$1 - P^+[S(\omega)] = \{1 - P^{\#}[S(\omega)]\}^{1/K},$$

so that

$$P^+[S(\omega)] = 1 - \exp\left\{ \sum_{i=1}^I \sum_{j=1}^J n_{ij} \ln(1 - q_{ij}) \right\}. \quad (4.5.13)$$

It can be shown that (Anderson and Trifuanc (1977))

$$P^+[S(\omega)] \geq P^*[S(\omega)] \geq P[S(\omega)]. \quad (4.5.14)$$

Equality occurs between P^+ and P^* only when there is just one source element, or when q_{ij} is identical for all source elements. Equality between P^* and P occurs only in the case where $q_{ij} = 0$ for all source elements.

4.6 Summary

The steps in the derivation of a uniform frequency dependent risk functional can be summarized as follows:

- (1) Specify the geometry of earthquake zones, by point, line, areal, dipping plane and volume sources. For each of these source zones, the estimated number of events of each earthquake size, $N(M_j)$, is defined. The uncertainties in the estimation of seismicity and maximum allowed earthquake size are also defined. This is done by studying previous seismicity, by insights obtained from geological studies and plate tectonics, and by statistical inference and scientific judgment.
- (2) Divide each source zone into small source elements, and assuming the epicenter of each event of size M_j is equally likely to occur any place in the source zone, distribute the seismicity to each element accordingly.
- (3) Specify a frequency-dependent description of the attenuation of strong-motion amplitudes in the region, plus a description of the distribution of the observed amplitudes $S(\omega)$ about the mean estimated $\hat{S}(\omega)$. From this, define the function $q_{ij}[S(\omega)]$ which gives the probability that $S(\omega)$ will be exceeded at the site for an event of size M_j in the i -th element of the source zone.
- (4) Using (4.2.5), calculate $N_E[S(\omega)]$, the expected number of times that $S(\omega)$ will be exceeded at the site from all source elements in all source zones and all allowed earthquake sizes. Then calculate $P[S(\omega)]$.

the probability that $S(\omega)$ will be exceeded at least once at the site in the given period, using (4.4.6), (4.5.9) or (4.5.13).

- (5) Using the method described in Section 4.1, derive the frequency dependent uniform risk functional from the functions $P[S(\omega)]$.

This completes the theoretical description of our Model of Seismic Risk.

5. THE COMPUTER PROGRAM

5.1 Introduction

A Fortran computer program, NEQRISK, has been written which evaluates the uniform risk spectra using the formulae developed in the previous sections. It has been written as a continuation of our effort to accurately describe seismic hazard. The original version of the program, EQRISK, was first written by Anderson (1978) using the theory and formulae developed by Anderson and Trifunac (1977).

EQRISK was first developed on a DEC computer with a TOPS10 operating system capable of running programs with large memory. Subsequently, it was modified to run on mini-computer Data General Eclipse S-130 with a program size of no more than 128 K-bytes. NEQRISK represents the most recent generation of this computer program currently also running on Data General Eclipse S-130 with an AOS operating system. It is able to use either the original scaling relations of FS, SV and PSV developed before 1980 and used by Anderson (1978), or the new scaling relations of FS and PSV developed recently (Trifunac and Lee, 1985b,c). NEQRISK is also able to include the uncertainties in the estimation of the seismicity of the various types of faults.

The program NEQRISK includes options to read in point, line, areal, dipping plane and volume sources. The original version of EQRISK in 1978 also had the option of reading in dipping plane sources, but there the program simply replaced each surface element by a corresponding dipping element and the epicentral distance by the corresponding hypocentral distance. It was mentioned then that this was done because of the lack of a more realistic method. The method outlined in Section 2.3 for dipping plane source is now used in NEQRISK. For a dipping plane source, option allows either the assumption that the epicentral distance should be used or the assumption that the closest point to the fault should be used to calculate the risk. Similarly, for the line source, the program automatically uses the closest point. For point, areal (diffused zones) and volume sources, however, the only option allowed is the use of epicentral distance.

Besides including the uncertainties in the estimation of seismicity to calculate the seismic risk spectra, one other major difference between "NEQRISK" and "EQRISK" is the use of efficient and accurate algorithms for interpolation and differentiation. Interpolation is used in constructing the seismic risk spectra of spectral amplitudes $S(\omega)$ versus frequency ω of a fixed probability level. It involves reading the probability of exceedance curves of $P[S(\omega)]$ versus $\log_{10}[S(\omega)]$ for various frequencies ω and interpolating the spectral amplitudes at a fixed probability level. The original program "EQRISK" simply uses a straight line interpolation scheme between points. The current program "NEQRISK" uses an interpolation scheme involving the use of low pass digital filters (Lee and Trifunac, 1984), as was done in data processing of strong-motion earthquake accelerograms. Such interpolation filters can be described in a variety of ways,

as discussed in Chapter II of Lee and Trifunac (1984) for example, using window designs, equiripple designs, smooth designs, etc. The Finite Impulse Response (FIR) nonrecursive interpolation filter turns out to be the most efficient. The filter used here is one in this form, designed by Oetken et al (1975), in which the minimization of mean square criteria is used.

Differentiation is used to calculate the probability density function from the Poisson probability of exceedance. The original program "EQRISK" uses the forward central difference method to do the differentiation. A higher order differentiation formula as derived from the smooth low pass filter (Lee and Trifunac, 1984) is currently used in the updated program "NEQRISK." This gives more accurate end results.

5.2 Input Data

The input data can be divided into three main parts:

- (a) Input Part 1: Site Information and Model Parameters,
- (b) Input Part 2: Spectrum Scaling Data, and
- (c) Input Part 3: Seismicity Information.

The following computer listing is a description of these three parts of the input.

 * SITES.DAT *

INPUT PART 1 MODEL IDENTIFICATION, SITE PARAMETERS

```

LINE 1 944(1412, 2X, F10.2) MTY, MRS, MAL, ILTL, IPPL, IMRAC, MRL, IPPC,
*****
      USING THE UPDATED FS & PSV STATISTICS DEVELOPED IN 1984-85:
      MTY = 0      TERMINATE PROGRAM
      1      FOURIER SPECTRUM; SEISMICITY MODEL USES
      MAGNITUDE STATISTICS, S=0,1,2 SITE CONDITIONS
      2      ALL RESPONSE SPECTRA; SEISMICITY MODEL USES
      MAGNITUDE STATISTICS, S=0,1,2 SITE CONDITIONS
      3      FOURIER SPECTRUM; SEISMICITY MODEL USES
      INTENSITY STATISTICS, S=0,1,2 SITE CONDITIONS
      4      ALL RESPONSE SPECTRA; SEISMICITY MODEL USES
      INTENSITY STATISTICS, S=0,1,2 SITE CONDITIONS
      5      FOURIER SPECTRUM; MAGNITUDE STATISTICS;
      SITE CONDITION IS DEPTH TO BASEMENT H
      6      RESPONSE SPECTRUM, PSV ONLY; MAGNITUDE STATISTICS
      SITE CONDITION IS DEPTH TO BASEMENT H
      7      FOURIER SPECTRUM; INTENSITY STATISTICS;
      SITE CONDITION IS DEPTH TO BASEMENT H
      8      RESPONSE SPECTRUM, PSV ONLY; INTENSITY STATISTICS
      SITE CONDITION IS DEPTH TO BASEMENT H.

      MRS = 0      FOURIER SPECTRUM
      1      PSV 0 PERCENT DAMPING
      2      PSV 2 PERCENT
      3      PSV 5 PERCENT
      4      PSV 10 PERCENT
      5      PSV 20 PERCENT

      MAL - IDENTIFIES THE ATTENUATION LAW
      FOR MTY=1, 2
      MAL=1      FREQUENCY DEPENDENT ATTENUATION FUNCTION (1985)
      ONLY OPTION
      FOR MTY=3, 4
      MAL=1      ANDERSON PARAMETERS FOR WEST USA INTENSITY DECAY
      2      ANDERSON PARAMETERS FOR EAST USA INTENSITY DECAY

      ILTL = 0      SUPPRESSES ALL 'LITERAL' CALCULATIONS.
      1      INCLUDES BOTH 'LITERAL' AND 'POISSON' OUTPUT
      2      SOME SOURCES ARE 'LITERAL', OTHERS ARE 'POISSON'
      IN THIS CASE, THE ENTIRE PART 3 OF THE INPUT
      IS REPEATED TWICE. THE FIRST TIME THROUGH,
      IT READS THE POISSON SOURCES. THE SECOND
      TIME IT READS THE LITERAL SOURCES. COMPUTE
      TIME MAY BE DOUBLED.

      IPPL = 0      NORMAL VALUE
      -1      PRODUCES PRINTER PLOT OF DERIVATIVE OF P(S(T))
      AT ALL PERIODS CALCULATED IN THE SPECTRUM.
      1-11      PRODUCES PRINTER PLOT OF DERIVATIVE OF P(S(T))
      AT THE PERIOD 1-11 CORRESPONDING TO INPUT.

      IMRAC = 0      NORMAL VALUE
      1      READS ADDITIONAL LINE TO MODIFY DEFAULT DISTANCE-
      AREA CRITERIA FOR DIFFUSE SEISMICITY ZONES.

      MRL = 0 OR 1  DIFFUSE SOURCES ARE TREATED AS POINT SOURCES.
      2-10      MAGNITUDE - RUPTURE LENGTH RELATION AS FOLLOWS
      FLNG= 10***(A*M-B)
      MRL.      A      B      SOURCE
      1      0.0      0.4
  
```

Listing 1.1

```

C      2      0.6      2.7      APPROX. PRESS
C      3      0.67     2.23     THATCHER & HANKS 1.7 BARS
C      4      0.67     1.41     THATCHER & HANKS 0.1 BARS
C      5      0.67     3.41     THATCHER & HANKS 100 BARS
C      6      0.53     1.47     WYSS & BRUNE
C      7      1.02     5.77     TOCHER
C      8      1.32     7.99     OKAMOTO
C      9      0.395    1.454    HOUSNER M.I.T. 6.4
C      10     0.900    4.673    HOUSNER M.GE. 6.4
C      10     1.596    7.56     DER-KIUREGHIAN
C                                     USES EXP(.), NOT 10**(. )
C
C      IPPC = 0  NORMAL VALUE
C      1  PRINTS FINAL DISTANCE AND SEISMICITY ARRAY.  THIS
C          CAN BE HELPFUL IN LEARNING THE SOURCE DISTRIBUTION,
C          AND THUS IN UNDERSTANDING THE RESULTS.
C
C      LSSUP = 0  NORMAL VALUE
C      1  SUPPRESSES LISTING OF SEISMICITY MODEL.
C
C      IDL1 = STARTING PERIOD NUMBER FOR DO LOOPS OVER PERIODS
C          DEFAULT IDL1=1.
C
C      IDL2 = ENDING PERIOD NUMBER FOR DO LOOPS OVER PERIODS
C          DEFAULT IDL2=11.
C
C      IDL3 = SKIP PARAMETER FOR DO LOOPS OVER PERIODS.
C          DEFAULT IDL3=1.
C          BY DEFAULT, THE PROGRAM FINDS SPECTRA AT 11 PERIODS.
C          IF THESE ARE NOT ALL NEEDED, THEN IDL1, IDL2, IDL3 ALLOW
C          A FLEXIBLE WAY TO SELECT JUST THOSE THAT ARE NEEDED.
C
C      ISTD = 0  DEFAULT AND NORMAL VALUE
C      1  STOPS PROGRAM AFTER INTEGRATION IS SET UP.
C          DO NOT USE WITH ILTL=2.
C
C      IMODE =0  DEFAULT, LONG & LAT INPUT IN DECIMALS
C      =1,      LONG & LAT INPUT IN DEG MIN & SEC
C
C      YRS = MULTIPLICATION FACTOR FOR SEISMICITY - DEFAULT = 1.0
C          ONLY USED FOR POISSONIAN SEISMICITY - LITERAL RATES
C          ARE ALWAYS EXACTLY AS INPUT
C
C      LINE 2 904(2F10.5, I5, F5.0, 2I5, 20A2) SLONG, SLAT, IV, HIS, NPEX, MUST,
C          ***** (NSITE(20))
C          SLONG = WEST LONGITUDE OF SITE
C          SLAT = NORTH LATITUDE OF SITE
C          IV = 0 HORIZONTAL
C              1 VERTICAL
C          HIS = 0 SOFT, ALLUVIAL SITE, FOR MTY=1,2,3,4
C              1 INTERMEDIATE SITE, FOR MTY=1,2,3,4
C              2 HARD SITE, FOR MTY=1,2,3,4
C              = DEPTH TO GEOLOGIC BASEMENT (KM.) FOR MTY=5,6,7,8
C          NPEX = NUMBER OF PROBABILITIES OF EXCEEDANCE FOR WHICH SPECTRA
C              WILL BE CALCULATED (LIMIT 24)
C          MUST = SITE NUMBER
C          NSITE = ALPHANUMERIC SITE NAME
C      LINE 3 938(8F10.3) (PE(I), I=1, NPEX)
C          *****
C          PE(I) = EXCEEDANCE PROBABILITIES FOR THE SPECTRA.  USE ONLY AS
C              MANY LINES AS NEEDED.
C      LINE 4 (WHEN INRAC=1) CONTROLS THE INTEGRATION OF DIFFUSE SOURCES.
C          *****
C      LINE 4 910(I1, F9.1, 7F10.1) NC, (RC(I), AC(I), I=1, NC)
C          NC = 1 TO 4
C          RC, AC ... NC PAIRS OF DISTANCE - AREA CRITERIA.  MUST HAVE
C              RC(I) LESS THAN RC(I-1) AND AC(I) LT AC(I-1).
C              WITH A DIFFUSE SOURCE, NONE OF THE ELEMENTS OF
C              AREA WITH A DISTANCE LESS THAN RC(I) FROM THE SITE
C              WILL HAVE AN AREA GREATER THAN AC(I) IN THE

```


II.59

C
C
C
C
C

INTEGRATION.
THIS LINE IS READ ONLY WHEN IMRAC=1.
OTHERWISE, REASONABLE DEFAULT VALUES ARE ASSIGNED.

Listing 1.3

126

BEST COPY AVAILABLE


```

*****
*          SPEC. DAT          *
*****

```

INPUT PART 2 MODEL PARAMETEPS

THE REGRESSION PARAMETERS FOR ALL THE SCALING RELATIONS ARE STORED IN THE DATA BLOCK 'SPEC. DAT'. THE INPUT PARAMETERS 'MTV, MRS, MAL' TELL THE PROGRAM WHICH PARTS OF THIS BLOCK TO READ.

THE USER DOES NOT NORMALLY NEED TO DO ANYTHING WITH THIS DATA BLOCK, EXCEPT TO MAKE IT AVAILABLE TO THE COMPUTER. ITS ORGANIZATION IS LISTED BELOW.

ATTENUATION CURVES

FREQUENCY DEPENDENT ATTENUATION FUNCTION FOR CALIFORNIA
(TRIFUNAC AND LEE, 1985A)

ANDERSON PARAMETERS FOR WESTERN U. S.

ANDERSON PARAMETERS FOR EASTERN U. S.

SCALING PARAMETERS

FOURIER SPECTRUM FROM MAG-SITE MODEL

RESPONSE SPECTRUM FROM MAG-SITE MODEL (5 DAMPINGS)

FOURIER SPECTRUM FROM MMI-SITE MODEL

RESPONSE SPECTRUM FROM MMI-SITE MODEL (5 DAMPINGS)

FOURIER SPECTRUM FROM MAG-DEPTH MODEL

RESPONSE SPECTRUM FROM MAG-DEPTH MODEL (5 DAMPINGS)

FOURIER SPECTRUM FROM MMI-DEPTH MODEL

RESPONSE SPECTRUM FROM MMI-DEPTH MODEL (5 DAMPINGS)

COMMENTS FOLLOW ON SPECIFIC PARTS.

RICHTER AMPLITUDE - DISTANCE CURVE.

LINE 1 913(4I2, 10A2) NAME(10)

NAME(10) IDENTIFICATION OF AMPLITUDE - DISTANCE CURVE.

LINE 2 912(I2, 2A2, 12F6.1) NPER, ID, PER(12)

NPER = NUMBER OF PERIODS

ID IDENTIFICATION OF ARRAY

PER ARRAY OF PERIODS

LINE 3 AOR ARRAY OF COEFFICIENTS FOR ATTENUATION FUNCTION

LINE 4 SZM ARRAY OF COEFFICIENTS FOR ATTENUATION FUNCTION

ANDERSON PARAMETERS FOR WESTERN U. S. A.

ANDERSON PARAMETERS FOR EASTERN U. S. A.

SCALING PARAMETERS

911(I2, 2A2, 12F6.3)NX, ID, X(12)

A BLOCK OF RELATED LINES, ALL ON THE SAME FORMAT.

NX = NUMBER OF POINTS IN ARRAY X. SAME ON ALL LINES.

ID IDENTIFICATION OF ARRAY. FOR USER CHECKING.

X ONE OF TEN ARRAYS.

A, B, C, D, E 5 STATISTICS COEFFICIENTS

SEE PAPER BY TRIFUNAC & LEE (1985B) / PRELIMINARY EMPIRICAL
MODEL FOR SCALING FOURIER AMPLITUDE SPECTRA ...

MU, ST MEAN AND STANDARD DEVIATION OF GAUSSIAN CURVE FIT

ALP, BET COEFFICIENTS FOR RAYLEIGH FIT OF PSV RESIDUALS

INPUT PART 3 MODEL OF SEISMICITY

LINE 1 905(515,5X,110) NPONT,NLINE,NDFUS,NPINE,NVOLI,MOSN

NVOLM = NUMBER OF DIPPING VOLUME SOURCES. THE FOCUS
HAS UNIFORM PROBABILITY OF OCCURRING ANYPLACE IN
THE DIPPING VOLUME.
MOSN = MODEL OF SEISMICITY NUMBER

POINT SOURCE. SUBROUTINE PSIN.

```

ISI      = SAME AS L1 FOR LINE SOURCE - EXPLAINED BELOW.
CONF     = SAME AS L1 (USED FOR PREDICTION CALCULATIONS).
JST      = 1      SOURCE IS A POINT
          2      EQ'S ARE UNILATERAL RUPTURE, RANDOM DIRECTION.
              RUPTURE LENGTH GIVEN BY MRL IN BLOCK 1, LINE 1.
SL       = NOT USED
NL       = NOT USED
NAME     = ALPHANUMERIC IDENTIFICATION OF THE SOURCE.

```

LINE P3 S903(2F10.5) XS,YS

LINE SOURCE. SUBROUTINE LSIN.

LINE L1 S901(I2, F2, 2, I2, F2, 0, I2, 35A2) ISI, CONF, JST, SL, NL, NAME(35)

Listing 3.1

II. 62

```

C      CONF = CONFIDENCE LEVEL FOR PREDICTED EARTHQUAKE. USED ONLY
C      WHEN ILTL=2. DEFAULT IS 0.0, WHICH MEANS THE
C      PREDICTION IS IGNORED UNLESS THIS PARAMETER IS SET.
C      THE PROCEEDURE ASSUMES THAT THERE IS ONLY ONE PREDICTED
C      EVENT, OR THAT THE CONFIDENCE IS THE SAME FOR ALL
C      PREDICTED EVENTS.
C      'CONF' NEGATIVE ALLOWS FINDING SPECTRA FOR MULTIPLE
C      CONFIDENCE LEVELS. THE ENTRY '-K' GIVES K+1 CONFIDENCE
C      LEVELS EVENLY SPACED BETWEEN 0.0 AND 1.0.
C      JST = 0-10 DESCRIBES THE MAGNITUDE - RUPTURE LENGTH RELATION.
C      SAME CHOICE AS FOR 'MRS' IN INPUT PART 1.
C      SL = STEP LENGTH (KM) FOR REPRESENTING THE LINE SOURCE.
C      DEFAULT = 5.0.
C      NL = NUMBER OF LINES TO INPUT SOURCE.
C      NAME = ALPHANUMERIC IDENTIFICATION OF FAULT.
C      LINE L2A (IF ISI=1,2,OR 3) (8F10.2) AAL,BBL,AMM,AMX,AZTM,AZTP
C      *****
C      AAL = COEFFICIENT 'A' IN LOG(N)=A-B*M (ISI=1,2)
C      BBL = COEFFICIENT 'B' IN LOG(N)=A-B*M (ISI=1,2,3)
C      AMM = MINIMUM MAGNITUDE (ISI=1,2,3; DEFAULT ALWAYS AVAILABLE)
C      AMX = MAXIMUM MAGNITUDE (ISI=1,2,3; DEFAULT ALWAYS AVAILABLE)
C      AZTM = USED TO DEFINE MOMENT RATE. (ISI=3)
C      AZTP = USED TO DEFINE MOMENT RATE. (ISI=3)
C      MOMENT RATE = AZTM * 10.0**AZTP
C      LINE L2B (IF ISI=4) (8F10.2) WS(12)
C      *****
C      WS(12)= SEISMICITY RATE ON THE FAULT.
C      REQUIRES TWO LINES WITH 12 VALUES.
C      BOTH MUST BE PRESENT EVEN IF ONE IS BLANK.
C      IF MAGNITUDE, RATE IS FOR M=3.0,3.5,4.0,...,8.0,8.5
C      IF INTENSITY, RATE IS FOR I=I,II,III,IV,...,X,XI,XII
C      LINES L3 S902 (2F10.5) XL(I),YL(I) --NL LINES--
C      *****
C      XL(I) = WEST LONGITUDE (DECIMAL DEGREES)
C      YL(I) = NORTH LATITUDE (DECIMAL DEGREES)
C      XL(I),YL(I), FOR I FROM 1 TO NL, GIVE THE COORDINATES
C      OF SUCCESSIVE POINTS ON THE FAULT.
C      DIFFUSE SOURCE. SUBROUTINE DSIN.
C      LINE D1 - SAME FORMAT AND VARIABLE NAMES AS L1 FOR LINE SOURCE.
C      *****
C      - USAGE OF FOLLOWING VARIABLES DIFFERS:
C      JST = 1 THE SOURCE RUPTURE LENGTHS ARE ZERO
C      = 2 THE SOURCES ARE ASSUMED TO BE UNILATERAL RUPTURE,
C      WITH THE EPICENTER HAVING EQUAL PROBABILITY TO BE
C      ANYPLACE IN THE ZONE. THE DIRECTION OF RUPTURE
C      IS RANDOM. THIS IS THE DEFAULT VALUE.
C      RUPTURE LENGTH GIVEN BY MRL IN BLOCK 1, LINE 1.
C      SL = NOT USED.
C      NL = NUMBER OF LONGITUDES USED TO DEFINE REGION.
C      LINE D2A OR D2B -- SAME AS L2A OR L2B FOR LINE SOURCE
C      *****
C      LINES D3 S902(3F10.5) YC(I),XL(I),XR(I) --NL LINES--
C      *****
C      ON EACH LINE, A NORTH LATITUDE 'YC(I)', AND THE
C      LONGITUDE AT WHICH IT INTERSECTS THE WEST 'XL(I)' AND
C      EAST 'XR(I)' BOUNDARY OF THE REGION.
C      'YC(I)' MUST INCREASE WITH SUCCESSIVE LINES.
C      THE FIRST VALUE OF 'YC(I)' IS THE SOUTHERN BOUNDARY OF THE
C      REGION, AND THE LAST VALUE IS THE NOUTHERN BOUNDARY.
C      DIPPING PLANAR SOURCE. SUBROUTINE DIPSIN.
C      SEE TEXT FOR THE CURRENT METHOD OF ASSIGNING PROBABILITY OF

```

II.63

```

C      OCCURRENCES IN THE SUB-REGIONS, DEPENDING OF THE MAGNITUDE
C      AND SIZE OF RUPTURE.
C      LINE DP1 - SAME AS LINE D1
C      *****
C      LINE DP2 - SAME AS LINE D2
C      *****
C      LINE DP3   S907(7F10.5)  PDPL(I), I=1,7
C      *****
C      PDPL(1),PDPL(2)  LONGITUDE AND LATITUDE OF A SURFACE POINT
C      TRACE OF THE DIPPING FAULT.
C      PDPL(3),PDPL(4)  LONGITUDE AND LATITUDE OF SECOND SURFACE POINT
C      ON THE SURFACE TRACE OF THE DIPPING PLANE.
C      PDPL(5),PDPL(6),PDPL(7)  LONGITUDE, LATITUDE, AND DEPTH(KM) OF
C      THIRD POINT ON THE DIPPING FAULT.
C      LINES DP4  --  SAME AS LINES D3
C      *****
C      VOLUME SOURCE SUBROUTINE VOLSIN
C      AT PRESENT, EXTENDED SOURCES CANNOT BE USED, ONLY POINT SOURCES ARE
C      TO BE USED.
C      INPUT FOR VOLUME SOURCE IS SAME AS THAT FOR DIFFUSE & DIPPING PLANE
C      SOURCE. THE SURFACE OF THE VOLUME IS DEFINED AS THAT OF THE
C      DIFFUSE SURFACE, AND ITS DEPTH DEFINED AS THAT IN LINE DP3 ABOVE.
C
C
C

```

5.3 The Flow Chart and Program Description

Figure 5.3.1 is a flow chart of the program NEQRISK. Each rectangular box corresponds to the calling of a subroutine whose name appears in the box. The subroutine usually will perform a specific task and subsequently return to the MAIN program. Figure 5.3.2 is an additional flow chart for the subroutine INSEIS. It involves calling 8 more subroutines which are not shown on the previous flow chart in Figure 5.3.1. For each of the subroutines whose names appear on these two figures, a brief description of its function will be presented.

II. 65

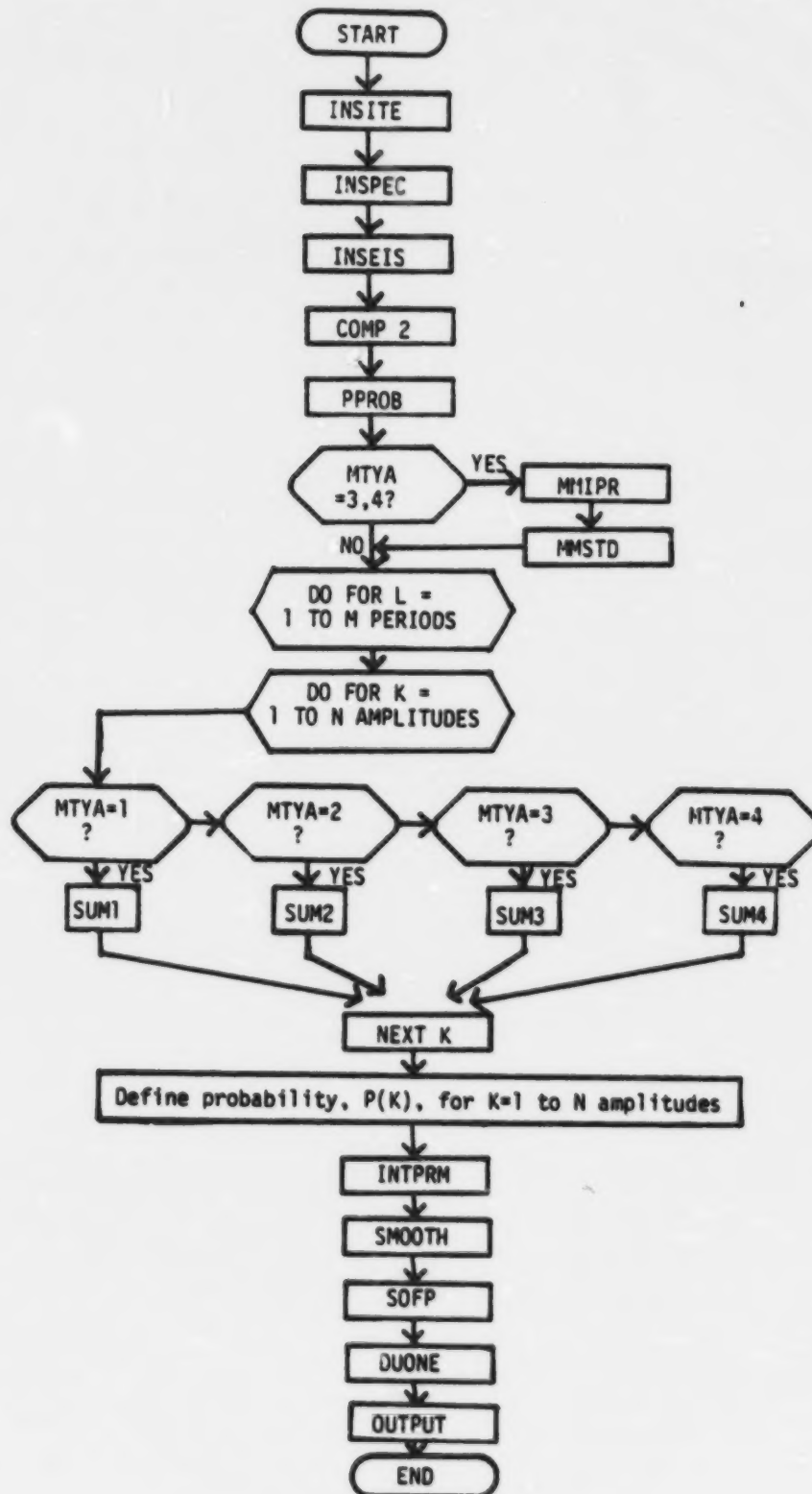


Figure 5.3.1 NEQRISK

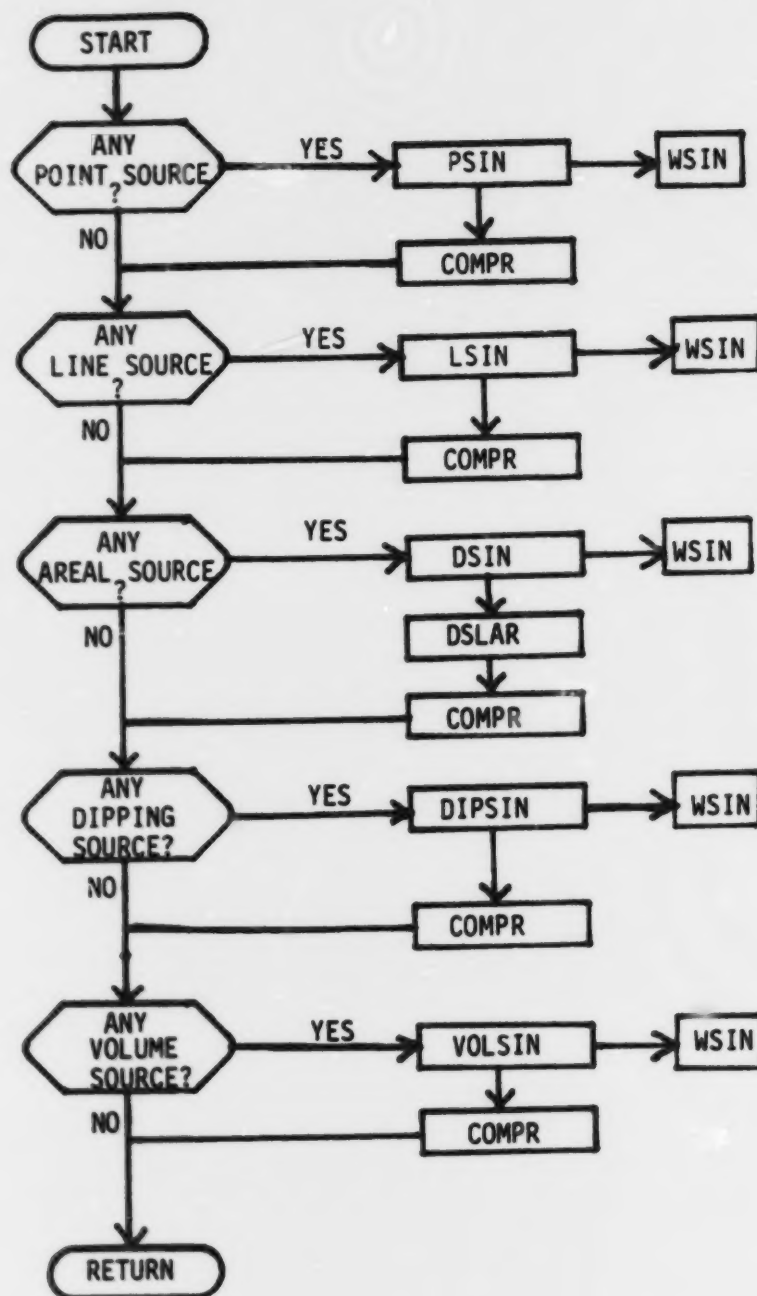


Figure 5.3.2 Subroutine INSEIS

5.3.1 INSITE

Subroutine INSITE reads in the model identification and site parameters from the data file SITE.DAT.

5.3.2 INSPEC

Subroutine INSPEC reads in the regression parameters for the particular scaling relations of the spectral amplitudes to be used in the analysis. For the old data base and the old scaling equations, this is stored in the data file SPEC.DAT. For the new data base and the updated scaling equations, this is stored in the data file ASPEC.DAT.

5.3.3 INSEIS

Subroutine INSEIS first reads in the number of various source types used. For each source type, through the corresponding subroutine, it reads in the necessary parameters for the calculation of seismicity at discrete earthquake sizes. It also reads in the maximum allowed earthquake size. In the new program NEQEISK, uncertainties associated with the estimation of seismicity parameters a and b and maximum allowed earthquake size are read in as confidence intervals of the corresponding parameters at a specified type of probability distribution. These are available in the file FSIN.DAT.

5.3.4 PSIN

This subroutine is called by subroutine INSEIS. It in turn calls WSIN to read in the parameters for the seismicity of a point source. It then calculates the expected number of earthquakes at each interval around discrete earthquake sizes.

5.3.5 WSIN

Subroutine WSIN is used to read in the seismicity parameters for all of the source types. It is called by PSIN, LSIN, DSIN and DPISIN. The seismicity parameters read in may include the uncertainties in the estimation and maximum allowed earthquake sizes. For each of the discrete earthquake sizes, WSIN computes the expected number of earthquakes of that size.

5.3.6 LSIN

Subroutine LSIN is called by subroutine INSEIS. It first calls WSIN to read in the parameters for the estimation of seismicity of line source. The line source is described by a set of straight lines, which are input as a consecutive sequence of coordinates as latitudes and longitudes. LSIN then represents the line source as a sequence of equally spaced points on the fault. The distances from these points to the observation point are calculated. The seismicity is then assigned to each point according to the probability of being the closest point that ruptures for each earthquake size.

5.3.7 DSIN

Subroutine DSIN is called by subroutine INSEIS. It first calls WSIN to read in the area boundaries and the parameters for the estimation of seismicity of the aeral source (diffuse zone). From the coordinates of the area boundary, DSIN generates a set of roughly square rectangles. Each of these squares may further be divided so that they are small enough, according to the criteria in the program, determined mainly by the epicentral distance between the site and the source elements. The seismicity is then divided among these areas in proportion to the area of each sub-element.

5.3.8 DIPSIN

Subroutine DIPSIN is called by INSEIS. It first calls WSIN to read in the area boundaries and the parameters necessary for the estimation of seismicity of the dipping plane source. As in subroutine DSIN, from the dimensions of the area boundary, DIPSIN generates a set of roughly square rectangles, which may further be divided so that they are small enough, according to the criteria of hypocentral distance and rupture area. Unlike DSIN, which divided the seismicity evenly among the elements, DIPSIN assigns the seismicity to each element according to the probability of being the closest element that ruptures for each earthquake size.

5.3.9 VOLSIN

Subroutine VOLSIN is called by INSEIS. It first calls WSIN to read in the volume boundaries and the parameters necessary for the estimation of seismicity of the volume source. VOLSIN then generates a set of roughly cubic elements, which may further be divided so that they are small enough, depending on the hypocentral distances to the site. As in DSIN, the seismicity is divided among the elements in proportion to the volume of each subelement. As pointed out earlier, this is only an approximate procedure for volume source representation and the procedure used here is only acceptable if the site is far enough. A criteria is set in the program which will automatically stop the program if the site is "too close" to the volume source.

5.3.10 COMPR

Subroutine COMPR is called by subroutines PSIN, LSIN, DSIN, DIPSIN and VOLSIN. Each time the seismicity is calculated and assigned to each of the source subelements of different source types, subroutine COMPR is

called to compress the seismicity and distance arrays so that all the points of equal distances are consolidated.

5.3.11 COMP2

Subroutine COMP2 is called by the MAIN program to compress the seismicity and distance arrays to eliminate any distances that have no seismicity.

5.3.12 PPROB

Subroutine PPROB is called by the MAIN program to compute the ratio $\gamma_{ij} = n_{ij}/N(M_j)$ defined in Section 4.3 for the calculation of probability of exceedance $P^+[S(\omega)]$ under the assumption that the number of earthquakes $N(M_j)$ is known (literal assumption).

5.3.13 MMIPR

Subroutine MMIPR is called by the MAIN program for the case of using the intensity (MMI) scaling functions to compute the intensity (MMI) at the site for a given magnitude and epicentral distance. An intensity attenuation scaling function for both eastern and western U.S. are available.

5.3.14 MMSTD

Subroutine MMSTD is called by the MAIN program for the case of using the intensity (MMI) scaling functions. It computes and prints the expected number of occurrences of each Modified Mercalli Intensity (MMI) at the site. The calculation is based on the attenuation function model developed by Anderson (1979b).

5.3.15 SUM1

Subroutine SUM1 is called by the MAIN program. Given a value of period, $PER(L)$, and a spectral amplitude, $SP(K)$, SUM1 performs the summation i from 1 to I over all source elements and the summation j from 1 to J over all earthquake sizes of $\bar{q}_{ij}E(n_{ij})$ to get $N_E(SP(K))$, the expected number of times that spectral amplitude $SP(K)$ of period $PER(L)$ will be exceeded at the site. The spectral amplitude assumed in the subroutine is the Fourier spectral amplitudes, FS , scaled in terms of magnitude, representative source-to-station distance and local site geology (Trifunac and Lee, 1985b).

5.3.16 SUM2

Subroutine SUM2 is called by the MAIN program. As in SUM1, given a value of period, $PER(L)$, and a spectral amplitude $SP(K)$, SUM2 performs the summation of $\bar{q}_{ij}E(n_{ij})$ to get $N_E(SP(K))$. The spectral amplitude assumed is pseudo relative velocity spectral amplitude, PSV , scaled in terms of magnitude, representative source-to-station distance and local site geology (Trifunac and Lee, 1985c).

5.3.17 SUM3

Subroutine SUM3 is called by the MAIN program. As in SUM1 and SUM2, given a value of period, $PER(L)$, and a spectral amplitude, $SP(K)$, SUM3 performs the summation of $\bar{q}_{ij}E(n_{ij})$ to get $N_E(SP(K))$. The spectral amplitude assumed is Fourier spectral amplitude, FS , scaled in terms of Modified Mercalli Intensity (MMI) and local site geology (using Trifunac and Lee, 1985b).

5.3.18 SUM4

Subroutine SUM4 is called by the MAIN program. As in SUM1, SUM2 and SUM3, given a value of period, $PER(L)$, and a spectral amplitude, $SP(K)$, SUM4 performs the summation of $\bar{q}_{ij}E(n_{ij})$ to get $N_E(SP(K))$. The spectral amplitude assumed is pseudo relative velocity response spectral amplitude, PSV, scaled in terms of Modified Mercalli Intensity (MMI) and local site geology (Trifunac and Lee, 1985c).

5.3.19 INTPRM

Subroutine INTPRM is called by the MAIN program. Having calculated the Poisson probability of exceedance for N spectral amplitudes at a particular period, INTPRM is called to interpolate the Poisson probabilities at intermediate amplitudes to result in 4 times as many points. INTPRM uses a finite impulse response (FIR) nonrecursive interpolation filter designed by Oetken et al (1975).

5.3.20 SMOOTH

Subroutine SMOOTH is called by the MAIN program. Having interpolated the Poisson probability of exceedance from N points to $4(N-1)+1$ points, SMOOTH is called to smooth the interpolated data using the $1/4$, $1/2$, $1/4$ weights. This is repeated several times.

5.3.21 SOFP

Subroutine SOFP is called by the MAIN program. It is called to construct a uniform risk spectra. For a given probability level, and for each discrete frequency ω , it calculates the spectral amplitude $S(\omega)$ that has the given probability that it will be exceeded. The probability versus spectral amplitudes $S(\omega)$ data to be used are the smoothed, interpolated set of data just calculated.

5.3.22 DUONE

Subroutine DUONE is called by the MAIN program. It is called to calculate the probability density function from the Poisson probability of exceedance. It uses a higher order differentiation formula derived from the smooth low-pass filter used in data processing of strong-motion accelerograms.

5.3.23 OUTPUT

Subroutine OUTPUT is called by the MAIN program. It prints out on a file the input site and model information, spectral information and seismicity information. For the logarithm of spectral amplitudes from -5 to +5, it prints out $N_E[S(\omega)]$, in the form of the logarithm of expected number of exceedances. In the same amplitude range, it prints out the probability of exceedances, the probability density function, and finally, the uniform risk spectra versus periods at a specified set of probability levels. It also plots out the expected number of exceedances versus periods and spectral amplitudes, the probabilities of exceedances, the probability density functions, both versus periods and spectral amplitudes, and finally the uniform risk spectra.

This completes the brief outline of the flow chart of the "NEQRISK" computer program.

5.4 Listing of Sample Inputs

5.4.1 Listing of Input Part 1

II. 75

DIRECTORY : UDD : EGRISK : WORKING : PEGRISK

SITES. DAT TXT 12-JUL-85 8:42:30 351

6 3 1 011	0 6 0 0 111	1 0	1.						
117. 90000	34. 10000	0 2. 3	17 1001	L. A. VICINITY	01				
0. 9	0. 5	0. 35	0. 2	0. 1	0. 05	0. 035	0. 02		
0. 01	0. 005	0. 0035	0. 0020	0. 0010	0. 00050	0. 00035	0. 00020		
0. 000100									

5.4.2 Listing of Input Part 2

 DIRECTORY UDD:EQRISK:WORKING:PEQRISK

SPEC.DAT UDF 12-JUL-85 8:44:24 17324

 0 0 1 3 FREQUENCY DEPENDENT ATTENUATION FUNCTION

12PER .040 .065 .11 .19 .34 .50 .90 1.6 2.8 4.4 7.5 14.0
 AOR-2.174-1.830-1.510-1.236-1.010 -.896 -.781 -.734 -.732 -.732 -.732 -.732
 SZM 29.27 29.39 29.51 29.65 29.78 29.87 30.01 30.14 30.28 30.38 30.51 30.66

0 0 1 2 WESTERN USA

12RMU 1.15 1.48 1.78 2.04 2.25 2.42 2.56 2.67 2.77 2.85 2.92 2.99
 RSG 0.3 0.3 0.3 0.3 0.3 0.3 0.3 0.3 0.3 0.3 0.3 0.3

0 0 2 2 EASTERN USA

12RMU 1.18 1.54 1.90 2.23 2.53 2.79 3.01 3.18 3.32 3.44 3.54 3.62
 RSG 0.3 0.3 0.3 0.3 0.3 0.3 0.3 0.3 0.3 0.3 0.3 0.3

1 0 0 8 FS AMPLITUDE MAG-SITE MODEL

12PER .040 .065 .11 .19 .34 .50 .90 1.60 2.80 4.40 7.50 14.00
 A -.258 -.019 .222 .433 .610 .706 .820 .883 .869 .712 .184 -.933
 B .041 .042 .033 -.003 -.057 -.084 -.102 -.110 -.122 -.121 -.086 .002
 C .086 .015 -.092 -.216 -.309 -.329 -.294 -.223 -.191 -.201 -.193 -.044
 D -1.373-1.681-2.207-2.965-3.844-4.394-5.100-5.487-5.395-4.741-2.924 .580
 E -.030 -.052 -.072 -.086 -.094 -.098 -.103 -.107 -.109 -.102 -.069 .012
 MU -.005 -.001 .002 .002 -.001 -.003 -.004 -.004 -.003 -.003 -.002 -.001
 ST .471 .460 .438 .402 .363 .351 .361 .388 .400 .388 .358 .331

2 1 0 9 PSV .00 DAMPING MAG-SITE MODEL

12PER .040 .065 .11 .19 .34 .50 .90 1.60 2.80 4.40 7.50 14.00
 A .010 -.008 .075 .271 .515 .666 .867 1.018 1.099 1.047 .712 -.083
 B -.004 .004 -.003 -.040 -.092 -.118 -.137 -.148 -.160 -.164 -.141 -.079
 C -.072 -.019 .002 -.026 -.092 -.136 -.166 -.139 -.107 -.104 -.100 -.077
 D -1.669-1.510-1.768-2.526-3.543-4.206-5.098-5.694-5.868-5.551-4.371-1.312
 E -.041 -.045 -.053 -.068 -.084 -.094 -.108 -.119 -.129 -.129 -.110 -.054
 ALP 1.247 1.252 1.220 1.211 1.256 1.290 1.268 1.166 1.211 1.530 2.225 2.086
 BET 1.026 1.006 1.005 1.007 1.000 .998 1.014 1.019 .935 .767 .492 .198
 N 10 10 10 10 10 10 10 10 9 6 3 2

2 2 0 9 PSV .02 DAMPING MAG-SITE MODEL

12PER .040 .065 .11 .19 .34 .50 .90 1.60 2.80 4.40 7.50 14.00
 A .002 .006 .123 .350 .577 .690 .832 .974 1.081 1.052 .758 .059
 B -.018 -.004 .000 -.025 -.069 -.093 -.115 -.131 -.150 -.158 -.138 -.075
 C -.099 -.056 -.033 -.046 -.100 -.140 -.171 -.145 -.112 -.110 -.115 -.087
 D -1.719-1.622-1.973-2.815-3.797-4.351-5.062-5.609-5.842-5.589-4.550-2.423
 E -.041 -.049 -.063 -.080 -.094 -.100 -.108 -.118 -.129 -.132 -.115 -.065
 ALP 1.314 1.328 1.281 1.239 1.263 1.299 1.287 1.185 1.220 1.543 2.271 3.209
 RET 1.026 1.007 1.007 1.009 1.002 1.000 1.016 1.022 .936 .767 .492 .200
 N 10 10 10 10 10 10 10 10 9 6 3 2

2 3 0 9 PSV .05 DAMPING MAG-SITE MODEL

12PER .040 .065 .11 .19 .34 .50 .90 1.60 2.80 4.40 7.50 14.00
 A .011 .018 .134 .363 .596 .699 .802 .906 1.013 1.020 .813 .260
 B -.021 -.009 -.007 -.026 -.063 -.085 -.105 -.121 -.142 -.151 -.135 -.074
 C -.101 -.069 -.051 -.059 -.104 -.144 -.178 -.154 -.118 -.112 -.118 -.102
 D -1.811-1.731-2.078-2.919-3.904-4.427-5.020-5.453-5.674-5.526-4.754-3.115
 E -.040 -.048 -.063 -.082 -.098 -.103 -.107 -.114 -.125 -.131 -.120 -.080
 ALP 1.316 1.350 1.313 1.263 1.267 1.292 1.276 1.187 1.245 1.586 2.327 3.262
 BET 1.026 1.007 1.008 1.010 1.002 1.001 1.017 1.023 .936 .768 .494 .203
 N 10 10 10 10 10 10 10 10 9 6 3 2

2 4 0 9 PSV .10 DAMPING MAG-SITE MODEL

12PER .040 .065 .11 .19 .34 .50 .90 1.60 2.80 4.40 7.50 14.00
 A .056 .039 .136 .355 .579 .674 .767 .878 1.002 1.029 .864 .410
 B -.027 -.016 -.012 -.028 -.061 -.080 -.100 -.117 -.140 -.150 -.135 -.075
 C -.103 -.076 -.059 -.066 -.108 -.146 -.182 -.163 -.127 -.117 -.123 -.118


```

D -1.974-1.970-2.186-2.993-3.934-4.419-4.966-5.407-5.674-5.583-4.954-3.632
E -.043 -.049 -.063 -.081 -.097 -.102 -.106 -.113 -.126 -.132 -.124 -.092
ALP 1.319 1.368 1.341 1.289 1.290 1.295 1.274 1.196 1.276 1.632 2.376 3.299
BET 1.026 1.007 1.000 1.010 1.003 1.002 1.018 1.023 .936 .769 .497 .212
N 10 10 10 10 10 10 10 10 10 9 6 3 2
2 5 0 9 PSV .20 DAMPING MAG-SITE MODEL
12PER .040 .065 .11 .19 .34 .50 .90 1.60 2.80 4.40 7.50 14.00
A .123 .092 .165 .348 .537 .616 .704 .820 .948 .986 .866 .512
B -.031 -.022 -.018 -.030 -.058 -.075 -.095 -.113 -.137 -.149 -.135 -.073
C -.096 -.084 -.076 -.081 -.111 -.142 -.176 -.165 -.130 -.114 -.118 -.135
D -2.218-2.116-2.389-3.098-3.922-4.354-4.864-5.297-5.567-5.511-5.017-3.982
E -.048 -.053 -.064 -.080 -.094 -.098 -.101 -.110 -.123 -.131 -.126 -.101
ALP 1.305 1.363 1.348 1.303 1.286 1.293 1.273 1.220 1.321 1.673 2.398 3.324
BET 1.026 1.008 1.009 1.010 1.003 1.002 1.019 1.024 .937 .769 .500 .219
N 10 10 10 10 10 10 10 10 10 9 6 3 2
3 0 0 7 FS AMPLITUDE MMI-SITE MODEL
12PER .040 .065 .11 .19 .34 .50 .90 1.60 2.80 4.40 7.50 14.00
A .104 .209 .253 .306 .352 .368 .369 .347 .310 .263 .179 .045
B .121 .101 .065 .019 -.021 -.036 -.044 -.056 -.079 -.081 -.029 .109
C .091 .028 -.090 -.227 -.323 -.334 -.277 -.200 -.157 -.148 -.144 -.143
D -2.118-1.774-1.476-1.397-1.540-1.659-1.709-1.542-1.225 -.911 -.519 -.061
MU -.003 .002 .006 .006 .004 .003 .005 .009 .010 .008 .002 -.006
ST .565 .516 .448 .394 .385 .405 .450 .467 .439 .397 .353 .329
4 1 0 8 PSV .00 DAMPING MMI-SITE MODEL
12PER .040 .065 .11 .19 .34 .50 .90 1.60 2.80 4.40 7.50 14.00
A .220 .229 .258 .305 .349 .364 .362 .337 .297 .257 .209 .155
B .093 .085 .060 .013 -.037 -.057 -.069 -.078 -.094 -.097 -.063 .029
C -.056 -.024 -.011 -.042 -.106 -.137 -.132 -.089 -.062 -.067 -.090 -.105
D -1.811-1.555-1.350-1.318-1.437-1.515-1.518-1.348-1.034 -.779 -.593 -.587
ALP 1.066 1.102 1.129 1.155 1.157 1.127 1.028 .945 1.078 1.457 2.168 2.999
BET 1.007 .988 .988 .990 .984 .982 .998 1.002 .917 .750 .483 .203
N 10 10 10 10 10 10 10 10 10 9 6 3 2
4 2 0 8 PSV .02 DAMPING MMI-SITE MODEL
12PER .040 .065 .11 .19 .34 .50 .90 1.60 2.80 4.40 7.50 14.00
A .207 .210 .239 .292 .347 .368 .372 .348 .309 .269 .220 .167
B .082 .085 .071 .036 -.008 -.029 -.044 -.058 -.079 -.087 -.059 .029
C -.090 -.062 -.044 -.063 -.118 -.149 -.147 -.105 -.074 -.078 -.104 -.122
D -1.854-1.625-1.465-1.488-1.650-1.744-1.752-1.562-1.235 -.957 -.746 -.715
ALP 1.141 1.163 1.169 1.188 1.206 1.193 1.107 1.012 1.118 1.490 2.231 3.139
BET 1.007 .989 .989 .990 .982 .981 .998 1.003 .916 .749 .482 .204
N 10 10 10 10 10 10 10 10 10 9 6 3 2
4 3 0 8 PSV .05 DAMPING MMI-SITE MODEL
12PER .040 .065 .11 .19 .34 .50 .90 1.60 2.80 4.40 7.50 14.00
A .209 .212 .236 .287 .341 .365 .372 .353 .316 .279 .231 .179
B .080 .081 .070 .040 .001 -.017 -.033 -.048 -.072 -.081 -.055 .028
C -.095 -.074 -.058 -.073 -.123 -.154 -.158 -.117 -.085 -.088 -.115 -.136
D -1.903-1.688-1.538-1.566-1.734-1.834-1.858-1.687-1.372-1.098 -.884 -.846
ALP 1.177 1.200 1.194 1.201 1.219 1.213 1.138 1.046 1.153 1.528 2.275 3.196
BET 1.006 .989 .989 .990 .982 .982 .999 1.004 .917 .749 .481 .203
N 10 10 10 10 10 10 10 10 10 9 6 3 2
4 4 0 8 PSV .10 DAMPING MMI-SITE MODEL
12PER .040 .065 .11 .19 .34 .50 .90 1.60 2.80 4.40 7.50 14.00
A .214 .216 .239 .285 .337 .360 .370 .354 .321 .287 .243 .195
B .076 .076 .065 .039 .005 -.012 -.027 -.043 -.066 -.075 -.052 .029
C -.096 -.083 -.070 -.083 -.128 -.156 -.163 -.127 -.092 -.094 -.122 -.148
D -1.958-1.768-1.621-1.649-1.809-1.906-1.936-1.784-1.495-1.240-1.040-1.000
ALP 1.212 1.236 1.223 1.218 1.232 1.227 1.158 1.079 1.201 1.580 2.309 3.188
BET 1.006 .989 .990 .991 .983 .983 1.001 1.006 .917 .748 .479 .199
N 10 10 10 10 10 10 10 10 10 9 6 3 2
4 5 0 8 PSV .20 DAMPING MMI-SITE MODEL
12PER .040 .065 .11 .19 .34 .50 .90 1.60 2.80 4.40 7.50 14.00
A .222 .224 .243 .284 .332 .354 .364 .350 .321 .292 .254 .211
B .068 .068 .059 .037 .006 -.010 -.025 -.040 -.061 -.069 -.048 .026

```

```

C - .099 - .089 - .081 - .093 - .131 - .158 - .166 - .134 - .101 - .099 - .124 - .150
D -2.027-1.856-1.735-1.758-1.902-1.991-2.015-1.871-1.611-1.389-1.214-1.171
ALP 1.253 1.280 1.265 1.248 1.250 1.240 1.172 1.114 1.271 1.662 2.359 3.148
BET 1.006 .990 .991 .992 .984 .983 1.002 1.006 .918 .748 .477 .195
N      10      10      10      10      10      10      10      10      9      6      3      2
5 0 0 8 FS AMPLITUDE MAG-DEPTH MODEL
12PER .040 .065 .11 .19 .34 .50 .90 1.60 2.80 4.40 7.50 14.00
A - .228 .014 .251 .432 .545 .594 .668 .734 .718 .549 .048 -.933
B .018 .014 .012 .017 .033 .045 .063 .082 .093 .089 .063 .013
C .077 .007 -.090 -.219 -.308 -.325 -.287 -.219 -.196 -.214 -.196 -.050
D -1.401-1.714-2.241-2.967-3.724-4.164-4.766-5.187-5.141-4.477-2.698 .556
E -.034 -.056 -.076 -.086 -.089 -.090 -.092 -.096 -.098 -.090 -.058 .011
MU -.004 .000 .003 .002 .001 .003 -.003 -.004 -.004 -.004 -.003 -.000
ST .473 .462 .438 .401 .364 .351 .353 .366 .367 .356 .338 .328
6 1 0 9 PSV .00 DAMPING MAG-DEPTH MODEL
12PER .040 .065 .11 .19 .34 .50 .90 1.60 2.80 4.40 7.50 14.00
A - .021 -.018 .070 .237 .408 .505 .651 .786 .839 .747 .407 -.276
B .039 .030 .027 .035 .052 .063 .076 .088 .097 .095 .079 .046
C -.076 -.024 -.002 -.029 -.090 -.130 -.153 -.125 -.100 -.108 -.121 -.091
D -1.500-1.465-1.747-2.478-3.345-3.877-4.627-5.205-5.341-4.924-3.693-1.442
E -.041 -.045 -.054 -.066 -.076 -.082 -.091 -.101 -.100 -.106 -.086 -.038
ALP 1.258 1.259 1.224 1.212 1.250 1.297 1.296 1.235 1.327 1.673 2.373 3.216
BET 1.027 1.006 1.006 1.007 1.000 .998 1.013 1.018 .935 .767 .492 .197
N      10      10      10      10      10      10      10      10      9      6      3      2
6 2 0 9 PSV .02 DAMPING MAG-DEPTH MODEL
12PER .040 .065 .11 .19 .34 .50 .90 1.60 2.80 4.40 7.50 14.00
A - .036 -.009 .121 .331 .497 .555 .638 .768 .852 .788 .479 -.097
B .039 .032 .028 .034 .047 .055 .066 .079 .093 .095 .083 .051
C -.104 -.061 -.037 -.049 -.097 -.134 -.160 -.135 -.110 -.115 -.127 -.102
D -1.638-1.571-1.954-2.786-3.645-4.067-4.632-5.186-5.400-5.046-3.931-2.066
E -.040 -.049 -.064 -.079 -.089 -.090 -.094 -.102 -.112 -.110 -.092 -.052
ALP 1.316 1.334 1.290 1.252 1.279 1.316 1.317 1.253 1.340 1.696 2.441 3.377
BET 1.027 1.000 1.000 1.010 1.002 .999 1.014 1.019 .935 .767 .492 .197
N      10      10      10      10      10      10      10      10      9      6      3      2
6 3 0 9 PSV .05 DAMPING MAG-DEPTH MODEL
12PER .040 .065 .11 .19 .34 .50 .90 1.60 2.80 4.40 7.50 14.00
A - .040 -.003 .131 .344 .517 .571 .628 .728 .807 .764 .538 .092
B .042 .037 .032 .034 .044 .052 .065 .078 .090 .091 .080 .053
C -.101 -.072 -.054 -.062 -.103 -.140 -.168 -.143 -.114 -.117 -.131 -.113
D -1.697-1.672-2.060-2.889-3.730-4.154-4.638-5.099-5.296-5.026-4.167-2.723
E -.038 -.049 -.065 -.082 -.092 -.094 -.094 -.101 -.109 -.110 -.097 -.068
ALP 1.320 1.360 1.324 1.273 1.282 1.315 1.321 1.263 1.350 1.731 2.520 3.526
BET 1.028 1.000 1.000 1.010 1.003 1.000 1.014 1.018 .934 .767 .494 .200
N      10      10      10      10      10      10      10      10      9      6      3      2
6 4 0 9 PSV .10 DAMPING MAG-DEPTH MODEL
12PER .040 .065 .11 .19 .34 .50 .90 1.60 2.80 4.40 7.50 14.00
A .007 .016 .128 .334 .500 .550 .603 .700 .787 .762 .581 .241
B .043 .038 .033 .035 .044 .052 .064 .076 .087 .089 .079 .049
C -.105 -.000 -.062 -.067 -.107 -.144 -.175 -.153 -.123 -.124 -.138 -.132
D -1.867-1.809-2.160-2.968-3.795-4.171-4.605-5.047-5.265-5.044-4.336-3.229
E -.041 -.049 -.064 -.081 -.092 -.094 -.094 -.100 -.109 -.111 -.102 -.079
ALP 1.326 1.382 1.356 1.302 1.296 1.310 1.312 1.250 1.376 1.772 2.573 3.558
BET 1.028 1.009 1.009 1.011 1.003 1.001 1.015 1.018 .932 .766 .496 .210
N      10      10      10      10      10      10      10      10      9      6      3      2
6 5 0 9 PSV .20 DAMPING MAG-DEPTH MODEL
12PER .040 .065 .11 .19 .34 .50 .90 1.60 2.80 4.40 7.50 14.00
A .092 .063 .136 .315 .473 .520 .555 .637 .732 .739 .616 .338
B .046 .042 .038 .039 .046 .053 .063 .074 .087 .091 .081 .049
C -.094 -.065 -.078 -.082 -.110 -.139 -.170 -.157 -.128 -.120 -.131 -.150
D -2.151-2.044-2.322-3.035-3.883-4.150-4.520-4.906-5.144-5.036-4.511-3.575
E -.040 -.052 -.063 -.079 -.089 -.091 -.091 -.096 -.106 -.111 -.105 -.088
ALP 1.307 1.377 1.368 1.325 1.312 1.321 1.307 1.277 1.432 1.848 2.646 3.618
BET 1.028 1.010 1.010 1.011 1.003 1.001 1.016 1.019 .933 .767 .500 .218

```

N	10	10	10	10	10	10	10	10	9	6	3	2
7 0 0 7	FS AMPLITUDE				MMI-DEPTH MODEL							
12PER	.040	.065	.11	.19	.34	.50	.90	1.60	2.80	4.40	7.50	14.00
A	.189	.209	.247	.296	.334	.345	.339	.320	.290	.249	.170	.040
B	-.038	-.024	-.003	.022	.047	.063	.081	.095	.099	.091	.050	-.009
C	.084	.022	-.095	-.231	-.326	-.335	-.279	-.209	-.177	-.170	-.163	-.156
D	-1.993	-1.637	-1.388	-1.352	-1.522	-1.637	-1.687	-1.579	-1.357	-1.081	-.626	.039
MU	-.003	.003	.006	.005	.003	.003	.004	.006	.000	.006	.001	-.005
ST	.566	.518	.450	.392	.378	.393	.427	.433	.401	.365	.336	.331
8 1 0 8	PSV .00 DAMPING				MMI-DEPTH MODEL							
12PER	.040	.065	.11	.19	.34	.50	.90	1.60	2.80	4.40	7.50	14.00
A	.223	.231	.257	.297	.332	.341	.335	.315	.283	.248	.201	.145
B	-.027	-.023	-.009	.017	.050	.069	.088	.096	.097	.089	.070	.036
C	-.061	-.029	-.015	-.045	-.109	-.139	-.135	-.099	-.082	-.089	-.107	-.115
D	-1.715	-1.464	-1.278	-1.284	-1.435	-1.524	-1.551	-1.434	-1.287	-.983	-.757	-.612
ALP	1.061	1.096	1.126	1.167	1.194	1.181	1.105	1.055	1.219	1.605	2.288	3.066
BET	1.006	.987	.988	.991	.984	.982	.990	1.004	.919	.752	.484	.204
N	10	10	10	10	10	10	10	10	9	6	3	2
8 2 0 8	PSV .02 DAMPING				MMI-DEPTH MODEL							
12PER	.040	.065	.11	.19	.34	.50	.90	1.60	2.80	4.40	7.50	14.00
A	.208	.213	.241	.289	.333	.348	.347	.328	.294	.259	.211	.158
B	-.023	-.026	-.020	.001	.033	.054	.076	.086	.090	.086	.070	.036
C	-.093	-.066	-.040	-.067	-.122	-.152	-.153	-.117	-.093	-.099	-.120	-.132
D	-1.766	-1.540	-1.393	-1.441	-1.625	-1.726	-1.756	-1.630	-1.383	-1.142	-.900	-.747
ALP	1.139	1.159	1.165	1.193	1.236	1.241	1.183	1.121	1.267	1.653	2.359	3.176
BET	1.006	.988	.988	.990	.983	.981	.990	1.004	.918	.751	.482	.201
N	10	10	10	10	10	10	10	10	9	6	3	2
8 3 0 8	PSV .05 DAMPING				MMI-DEPTH MODEL							
12PER	.040	.065	.11	.19	.34	.50	.90	1.60	2.80	4.40	7.50	14.00
A	.213	.215	.238	.283	.328	.344	.347	.333	.305	.271	.223	.168
B	-.024	-.025	-.019	.000	.031	.050	.071	.083	.088	.086	.072	.041
C	-.101	-.080	-.064	-.077	-.126	-.155	-.163	-.134	-.112	-.116	-.135	-.148
D	-1.824	-1.607	-1.466	-1.513	-1.694	-1.790	-1.845	-1.746	-1.521	-1.286	-1.040	-.876
ALP	1.176	1.194	1.186	1.202	1.249	1.264	1.217	1.152	1.295	1.688	2.413	3.249
BET	1.005	.988	.989	.990	.982	.981	.999	1.005	.918	.750	.480	.199
N	10	10	10	10	10	10	10	10	9	6	3	2
8 4 0 8	PSV .10 DAMPING				MMI-DEPTH MODEL							
12PER	.040	.065	.11	.19	.34	.50	.90	1.60	2.80	4.40	7.50	14.00
A	.219	.219	.239	.281	.324	.340	.345	.333	.300	.270	.235	.185
B	-.025	-.023	-.017	-.001	.027	.046	.069	.082	.086	.083	.070	.043
C	-.102	-.087	-.075	-.086	-.129	-.157	-.165	-.137	-.114	-.117	-.139	-.150
D	-1.882	-1.682	-1.551	-1.597	-1.769	-1.869	-1.921	-1.838	-1.634	-1.417	-1.190	-.939
ALP	1.211	1.231	1.215	1.216	1.256	1.276	1.244	1.195	1.341	1.725	2.425	3.231
BET	1.004	.988	.989	.991	.983	.982	1.000	1.006	.919	.749	.477	.192
N	10	10	10	10	10	10	10	10	9	6	3	2
8 5 0 8	PSV .20 DAMPING				MMI-DEPTH MODEL							
12PER	.040	.065	.11	.19	.34	.50	.90	1.60	2.80	4.40	7.50	14.00
A	.226	.228	.245	.280	.319	.335	.343	.331	.300	.282	.245	.199
B	-.022	-.024	-.010	-.000	.027	.043	.062	.073	.079	.079	.068	.043
C	-.102	-.093	-.086	-.096	-.132	-.157	-.168	-.143	-.117	-.117	-.140	-.161
D	-1.960	-1.791	-1.675	-1.700	-1.862	-1.956	-2.004	-1.920	-1.735	-1.550	-1.355	-1.205
ALP	1.251	1.277	1.259	1.250	1.277	1.290	1.254	1.220	1.394	1.787	2.458	3.190
BET	1.004	.988	.990	.991	.983	.982	1.000	1.006	.918	.748	.473	.185
N	10	10	10	10	10	10	10	10	9	6	3	2

5.4.3 Listing of Input Part 3

 DIRECTORY :UDD:EGRISK:WORKING:PEGRISK

FSIN.DAT TXT 12-JUL-85 8:46:56 1638

0	2	1	0	8				
5	0	2	511	SIERRA MADRE - CUCAMONGA FAULT ZONE				
			3.1482	0.86	2.75	7.5	3.8	24.0 2 2
			.25	0.1	1.			
			118.480	34.281	WEST END - 1971 BREAK			
			118.420	34.293	CENTER, 1971 BREAK			
			118.294	34.266	EAST END, 1971 BREAK			
			118.128	34.185	ALTADENA			
			118.000	34.158	RAYMOND FAULT			
			117.910	34.144	AZUSA			
			117.844	34.152	GLENDORA			
			117.735	34.118	CLAREMONT			
			117.645	34.163	ALTA LOMA			
			117.488	34.165	RIALTO-COLTON FAULT			
			117.425	34.196	SAN JACINTO FAULT - EAST END			
5	0	2	517	SAN ANDREAS FAULT - CAJON PASS TO SAN LUIS OBISPO				
			4.4660	0.86	2.75	8.5	7.9	25.0 2 2
			.25	0.1	1.			
			117.482	34.277	CAJON JCT.			
			117.627	34.345	WRIGHTWOOD			
			117.844	34.438	VALLERMO			
			118.285	34.612	LEONA VALLEY			
			118.713	34.759	SANDBERG			
			118.946	34.818	FRAZIER PARK			
			119.008	34.816	BIG PINE FLT. JCT			
			119.251	34.871	BIG BEND			
			119.364	34.911	BIG BEND			
			119.443	34.963	RT. 166 & 33			
			119.675	35.136	SOUTH CARRIZO PLAIN			
			119.885	35.329	HWY. 58			
			120.296	35.718	CHOLAME			
			120.337	35.806				
			120.424	35.889	PARKFIELD			
			121.248	36.647	"ABOVE S"			
			121.533	36.843	SAN JUAN BAUTISTA - S.END, 1906			
4	0	2	0 2	TRANSVERSE RANGES				
			0.0	9.6	2.8	0.84	0.25	0.073 0.022 .0064
			0.0	0.0	0.0	0.0	2 2	
			.25	0.1	1.			
			33.0	121.0	117.5			
			35.5	121.0	117.5			

Input 3.1

BEST COPY AVAILABLE

149

5.5 Listing of Printer Output

The following is a sample of a listing of the output data file from running "NEQRISK": RISK.OUTPUT.DAT. The data file consists of 5 parts:

(1) Listing of input site and model parameters, spectral data and seismicity information, as read in respectively from input data files "SITES.DAT," "SPEC.DAT" and "FSIN.DAT," which were listed in the previous Section 5.4.

(2) Listing of the output data, $N_E[S(\omega)]$, the expected number of exceedances on logarithmic scale, as a function of periods and spectral amplitudes. The discrete periods range from .04 sec. to about 2 sec., while the spectral amplitudes, in the form of $\log_{10}[S(\omega)]$, range from -5.00 to +5.00 in steps of .2.

(3) Listing of the output data, $P[S(\omega)]$, the probability of exceeding $S(\omega)$ at the site, again as a function of periods and spectral amplitudes in the same range.

(4) Listing of the output data $dP[S(\omega)/S(\omega)]$, the probability density function in the same range of periods and spectral amplitudes.

(5) Listing of the output uniform risk spectrum, which gives the $\log_{10}[S(\omega)]$ amplitudes in the same period range for a discrete set of probability levels.

Alternatively, the user can choose to have NEQRISK output only the the expected M.M.I. values, to be exceeded at specified probability levels. In such a case, only part (1) followed by a listing of such M.M.I. values will be the output data.

DIRECTORY :UDD:EQRISK:WORKING:PEQRISK

RISK_OUTPUT.DAT TXT 12-JUL-85 16:30:06 48593

THE FREQUENCY DEPENDENT SEISMIC RISK

THIS PROGRAM WAS ORIGINALLY DEVELOPED BY J. G. ANDERSON & M. D. TRIFUNAC,
SUBSEQUENTLY IT WAS UPDATED BY V. W. LEE & M. D. TRIFUNAC,
AND CURRENTLY RUNNING ON AOS ECLIPSE 5-130

CONTROL PARAMETERS

MTV	MRS	MAL	ILTL	IPPL	IMRAC	MRL	IPPC	ISSUP	IDL1	IDL2	IDL3	ISTP
6	3	1	0	11	0	6	0	0	1	11	1	0

SEE PROGRAM LISTING FOR EXPLANATION

YRS = 1.000

INPUT SEISMICITY RATES ARE MULTIPLIED BY YRS FOR RISK CALCULATIONS

12PER	.040	.065	.110	.190	.340	.500	.900	1.600	2.800	4.400	7.500	14.000
0A0R-2	1.74	-1.830	-1.510	-1.236	-1.010	-.896	-.781	-.734	-.732	-.732	-.732	-.732
05ZM29	27029	39029	51029	65029	78029	87030	01030	14030	20030	30030	51030	660
12PER	.040	.065	.110	.190	.340	.500	.900	1.600	2.800	4.400	7.500	14.000
0 A	-.040	-.003	.131	.344	.517	.571	.628	.728	.807	.764	.538	.092
0 B	.042	.037	.032	.034	.044	.052	.065	.078	.090	.091	.080	.053
0 C	-.101	-.072	-.054	-.062	-.103	-.140	-.168	-.143	-.114	-.117	-.131	-.113
0 D	-1.697	-1.672	-2.060	-2.889	-3.750	-4.154	-4.638	-5.099	-5.296	-5.026	-4.167	-2.723
0 E	-.038	-.049	-.065	-.082	-.092	-.094	-.094	-.101	-.109	-.110	-.097	-.068
0ALP	1.320	1.360	1.324	1.273	1.202	1.315	1.321	1.263	1.358	1.731	2.520	3.526
0BET	1.020	1.000	1.000	1.010	1.003	1.000	1.014	1.018	.934	.767	.494	.200
0 N	10	10	10	10	10	10	10	10	9	6	3	2

THE RISK IS FOUND FOR THE SITE

L. A. VICINITY 01

SITE NO. 1001

117.90000 DEG WEST LONGITUDE

34.10001 DEG NORTH LATITUDE

IV = 0

HIS = 2.300

RISK FOUND FOR 17 PROBABILITIES OF EXCEEDANCE --

.90000	.50000	.35000	.20000	.10000
.05000	.03500	.02000	.01000	.00500
.00350	.00200	.00100	.00050	.00035
.00020	.00010			

1 THE FOLLOWING SOURCES TREATED AS POISSON
MODEL OF SEISMICITY NO. (NOSH)... 0

THE SEISMICITY IS REGARDED AS A SUPERPOSITION OF

- 0 POINT SOURCES
- 2 LINE SOURCES (FAULTS)
- 1 REGIONS OF DIFFUSE SEISMICITY
- 0 DIPPING PLANES
- 0 DIPPING VOLUMES

LINE SOURCE SIERRA MADRE - CUCAMONGA FAULT ZONE
SEISMICITY INPUT BY MOMENT RATE (PER YEAR), B, MMAX

A, B, M-MIN, M-MAX, MOMENT/YR

31482D 01 . 860000 00 2.750 7.500 .380000 25

SCALED UP BY 1 YEARS.

LPTYPE, MPTYPE, DAAL, DBBL, DMXM 2 2 .25 .10 1.00

LOCATIONS OF ENDS ARE

118.48000 WEST LONGITUDE 34.28101 NORTH LATITUDE

117.42500 WEST LONGITUDE 34.19600 NORTH LATITUDE

FAULT IS REPRESENTED BY 11 - 1 STRAIGHT LINE SEGMENTS

ISI= 3

SEISMICITY	# OF EARTHQUAKES	ASSUMED FAULT LENGTH
MAG 3.0	2.1980240	.1259
3.5	.8378583	.2512
4.0	.3194134	.5012
4.5	.1221269	1.0000
5.0	.0467860	1.9953
5.5	.0179578	3.9811
6.0	.0069857	7.9432
6.5	.0026685	15.8489
7.0	.0008985	31.6226
7.5	.0001985	63.0954
8.0	.0000192	105.0000
8.5	.0000000	105.0000

THE FAULT IS REPRESENTED BY 22 POINTS AT SPACING OF 5.00 KILOMETER

THIS GIVES A LENGTH OF 105.00 KILOMETER

INSERTED IN DISTANCE AND SEISMICITY ARRAYS IN POSITIONS 1 TO 22

LINE SOURCE SAN ANDREAS FAULT - CAJON PASS TO SAN LUIS OBISPO

SEISMICITY INPUT BY MOMENT RATE (PER YEAR), B, MMAX

A, B, M-MIN, M-MAX, MOMENT/YR

44660D 01 . 860000 00 2.750 8.500 .790000 26

SCALED UP BY 1 YEARS.

LPTYPE, MPTYPE, DAAL, DBBL, DMXM 2 2 .25 .10 1.00

LOCATIONS OF ENDS ARE

117.48200 WEST LONGITUDE 34.27699 NORTH LATITUDE

121.53300 WEST LONGITUDE 36.84300 NORTH LATITUDE

FAULT IS REPRESENTED BY 17 - 1 STRAIGHT LINE SEGMENTS

ISI= 3

SEISMICITY	# OF EARTHQUAKES	ASSUMED FAULT LENGTH
MAG 3.0	45.6957600	.1259
3.5	17.4819900	.2512
4.0	6.6484300	.5012
4.5	2.5389540	1.0000
5.0	.9726570	1.9953
5.5	.3733327	3.9811
6.0	.1435649	7.9432
6.5	.0553897	15.8489
7.0	.0213471	31.6226
7.5	.0082536	63.0954
8.0	.0027971	125.0919
8.5	.0006201	251.1867

THE FAULT IS REPRESENTED BY 96 POINTS AT SPACING OF 5.00 KILOMETER

THIS GIVES A LENGTH OF 475.00 KILOMETER

INSERTED IN DISTANCE AND SEISMICITY ARRAYS IN POSITIONS 1 TO 96

DIFFUSE SEISMICITY REGION TRANSVERSE RANGES

N(M) READ DIRECTLY

SEISMICITY # OF

```

MAG    EARTHQUAKES
3.0      0000000
3.5      11.2274700
4.0      3.3633060
4.5      1.0383550
5.0      .3186481
5.5      .0961237
6.0      .0299835
6.5      .0090447
7.0      .0000000
7.5      .0000000
8.0      .0000000
8.5      .0000000

```

EVENTS ARE ASSUMED TO BE UNILATERAL RUPTURES

INPUT GIVES LONGITUDE OF EAST AND WEST BOUNDARY AT 2 LATITUDES
INITIALLY 4 RECTANGLES

REGION HAS AREA 89388.1

BEGINNING AREA-DISTANCE CHECKS

AREA, CORNER DISTANCE, CENTER DISTANCE

10000.000 220.000 339.052

NOW 16 RECTANGLES

REGION HAS AREA 89388.1

AREA, CORNER DISTANCE, CENTER DISTANCE

1140.000 75.000 154.057

NOW 80 RECTANGLES

REGION HAS AREA 89388.1

AREA, CORNER DISTANCE, CENTER DISTANCE

130.000 25.000 51.693

NOW 184 RECTANGLES

REGION HAS AREA 89388.1

AREA, CORNER DISTANCE, CENTER DISTANCE

15.000 9.000 18.014

NOW 304 RECTANGLES

REGION HAS AREA 89388.1

INSERTED INTO RA, SA ARRAYS IN ELEMENTS 1 TO 304

ML, MX 1 12

MAGNITUDE-RUPTURE LENGTH NO. . . 6

3.0 .1259

3.5 .2512

4.0 .5012

4.5 1.0000

5.0 1.9953

5.5 3.9811

6.0 7.9432

6.5 15.8489

7.0 31.6226

7.5 63.0954

8.0 125.8919

8.5 251.1867

DATA COMPRESSED TO 144 DISTANCES FOR INTEGRATION

LOG10(EXPECTED NUMBER OF EXCEEDANCES) --- IV = 0 HIS = 2.300

LOG(SL)	04000	06500	11000	19000	34000	50000	90000	1.60000
-5.00	1.93383	1.97000	1.97075	1.97076	1.97076	1.97076	1.97076	1.97076
-4.80	1.91010	1.96794	1.97070	1.97075	1.97076	1.97076	1.97075	1.97073
-4.60	1.88051	1.96260	1.97045	1.97068	1.97071	1.97072	1.97070	1.97059
-4.40	1.84516	1.95188	1.96945	1.97037	1.97051	1.97052	1.97046	1.97004
-4.20	1.80329	1.93456	1.96648	1.96926	1.96971	1.96973	1.96950	1.96829
-4.00	1.75383	1.91066	1.95960	1.96615	1.96733	1.96730	1.96666	1.96389
-3.80	1.69631	1.88077	1.94702	1.95930	1.96171	1.96152	1.96015	1.95514
-3.60	1.63126	1.84493	1.92795	1.94714	1.95122	1.95074	1.94838	1.94089
-3.40	1.55978	1.80206	1.90264	1.92903	1.93504	1.93423	1.93078	1.92098
-3.20	1.48252	1.75044	1.87151	1.90514	1.91325	1.91220	1.90768	1.89580
-3.00	1.39791	1.68890	1.83421	1.87577	1.88623	1.88508	1.87951	1.86554
-2.80	1.30097	1.61729	1.78923	1.84059	1.85393	1.85278	1.84607	1.82950
-2.60	1.18486	1.53587	1.73447	1.79817	1.81523	1.81412	1.80591	1.78571
-2.40	1.04609	1.44360	1.66811	1.74615	1.76786	1.76666	1.75632	1.73107
-2.20	.88875	1.33615	1.58859	1.68181	1.70868	1.70696	1.69360	1.66149
-2.00	.72223	1.20698	1.49356	1.60202	1.63387	1.63078	1.61309	1.57232
-1.80	.55500	1.05285	1.37880	1.50293	1.53916	1.53352	1.51001	1.46006
-1.60	.39093	.87830	1.23971	1.38086	1.42139	1.41255	1.38236	1.32483
-1.40	.22922	.69268	1.07560	1.23497	1.28089	1.26937	1.23261	1.17018
-1.20	.06581	.50350	.89121	1.06816	1.12089	1.10760	1.06509	1.00043
-1.00	-.10617	.31301	.69284	.88395	.94387	.92922	.88316	.81908
-.80	-.29710	.11830	.48390	.68338	.75031	.73499	.68925	.62830
-.60	-.51968	-.08798	.26351	.46564	.54046	.52607	.48467	.42849
-.40	-.78377	-.31722	.02711	.22945	.31485	.30326	.26936	.21896
-.20	-1.09142	-.58188	-.23235	-.02678	.07313	.06597	.04234	-.00153
-.00	-1.43522	-.89028	-.52218	-.30579	-.18668	-.18771	-.19794	-.23471
.20	-1.80106	-1.24381	-.84828	-.61183	-.46768	-.46038	-.45360	-.48281
.40	-2.17800	-1.63717	-1.21497	-.95016	-.77368	-.75555	-.72756	-.74881
.60	-2.57238	-2.06316	-1.62533	-1.32626	-1.10938	-1.07794	-1.02377	-1.03668
.80	-3.01145	-2.52411	-2.08316	-1.74542	-1.48078	-1.43415	-1.34780	-1.35139
1.00	-3.53084	-3.04149	-2.59708	-2.21301	-1.89558	-1.83318	-1.70767	-1.70265
1.20	-4.16350	-3.65104	-3.18425	-2.73633	-2.36311	-2.28662	-2.11475	-2.09963
1.40	-4.93866	-4.39232	-3.87006	-3.32786	-2.89470	-2.80833	-2.58398	-2.55694
1.60	-5.88234	-5.30592	-4.68569	-4.00812	-3.50477	-3.41425	-3.13294	-3.09091
1.80	-7.01540	-6.43307	-5.66700	-4.80757	-4.21279	-4.12322	-3.78114	-3.71961
2.00	-8.36159	-7.81581	-6.85652	-5.76756	-5.04622	-4.95951	-4.55034	-4.46329
2.20	-10.00000	-9.51042	-8.31011	-6.94019	-6.04479	-5.95784	-5.46651	-5.34574
2.40	-10.00000	-10.00000	-10.00000	-8.38869	-7.26378	-7.17018	-6.56423	-6.39682
2.60	-10.00000	-10.00000	-10.00000	-10.00000	-8.77574	-8.67077	-7.89578	-7.65851
2.80	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-9.54095	-9.19443
3.00	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000
3.20	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000
3.40	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000
3.60	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000
3.80	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000
4.00	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000
4.20	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000
4.40	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000
4.60	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000
4.80	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000
5.00	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000	-10.00000

POISSON PROBABILITY OF EXCEEDANCE --- IV= 0 HIS= 2.300									
LOG(SL)	PERIOD								
	04000	06500	11000	19000	34000	50000	90000	1. 60000	
-5.00	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	
-4.80	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	
-4.60	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	
-4.40	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	
-4.20	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	
-4.00	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	
-3.80	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	
-3.60	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	
-3.40	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	
-3.20	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	
-3.00	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	
-2.80	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	
-2.60	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	
-2.40	.99999	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	
-2.20	.99957	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	
-2.00	.99488	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	
-1.80	.97238	.99999	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	
-1.60	.91456	.99948	1.00000	1.00000	1.00000	1.00000	1.00000	1.00000	
-1.40	.81644	.99276	.99999	1.00000	1.00000	1.00000	1.00000	1.00000	
-1.20	.68765	.95874	.99958	.99999	1.00000	1.00000	.99999	.99996	
-1.00	.54382	.87203	.99277	.99953	.99985	.99980	.99952	.99863	
-.80	.39622	.73101	.95251	.99196	.99640	.99563	.99247	.98572	
-.60	.26082	.55808	.84030	.94616	.96892	.96519	.95276	.93159	
-.40	.15170	.38227	.65507	.81660	.87314	.86605	.84422	.80903	
-.20	.07782	.23040	.44327	.60945	.69376	.68777	.66793	.63082	
-.00	.03684	.12080	.25954	.39016	.47827	.47747	.46951	.44149	
.20	.01569	.05545	.13222	.21685	.28870	.29279	.29663	.28036	
.40	.00662	.02279	.05914	.10610	.15498	.16102	.17077	.16332	
.60	.00267	.00861	.02342	.04608	.07479	.08017	.09033	.08780	
.80	.00097	.00299	.00822	.01781	.03251	.03613	.04390	.04349	
1.00	.00029	.00091	.00253	.00610	.01264	.01458	.01941	.01963	
1.20	.00007	.00022	.00065	.00183	.00433	.00516	.00765	.00792	
1.40	.00001	.00004	.00013	.00047	.00127	.00155	.00260	.00277	
1.60	.00000	.00000	.00002	.00010	.00031	.00039	.00074	.00081	
1.80	.00000	.00000	.00000	.00002	.00006	.00008	.00017	.00019	
2.00	.00000	.00000	.00000	.00000	.00001	.00001	.00003	.00003	
2.20	.00000	.00000	.00000	.00000	.00000	.00000	.00000	.00000	
2.40	.00000	.00000	.00000	.00000	.00000	.00000	.00000	.00000	
2.60	.00000	.00000	.00000	.00000	.00000	.00000	.00000	.00000	
2.80	.00000	.00000	.00000	.00000	.00000	.00000	.00000	.00000	
3.00	.00000	.00000	.00000	.00000	.00000	.00000	.00000	.00000	
3.20	.00000	.00000	.00000	.00000	.00000	.00000	.00000	.00000	
3.40	.00000	.00000	.00000	.00000	.00000	.00000	.00000	.00000	
3.60	.00000	.00000	.00000	.00000	.00000	.00000	.00000	.00000	
3.80	.00000	.00000	.00000	.00000	.00000	.00000	.00000	.00000	
4.00	.00000	.00000	.00000	.00000	.00000	.00000	.00000	.00000	
4.20	.00000	.00000	.00000	.00000	.00000	.00000	.00000	.00000	
4.40	.00000	.00000	.00000	.00000	.00000	.00000	.00000	.00000	
4.60	.00000	.00000	.00000	.00000	.00000	.00000	.00000	.00000	
4.80	.00000	.00000	.00000	.00000	.00000	.00000	.00000	.00000	
5.00	.00000	.00000	.00000	.00000	.00000	.00000	.00000	.00000	

ABOVE DATA ARE THEN INTERPOLATED AND SMOOTHED...

FOLLOWING SPECTRA ASSUME EARTHQUAKES ARE POISSON, WITH MEAN AS INPUT
 ROBABILITY DENSITY FUNCTION
 LOG(SL)

	04000	06500	11000	19000	34000	50000	90000	1.50000
-5.00	00000	00000	00000	00000	00000	00000	00000	00000
-4.80	00000	00000	00000	00000	00000	00000	00000	00000
-4.60	00000	00000	00000	00000	00000	00000	00000	00000
-4.40	00000	00000	00000	00000	00000	00000	00000	00000
-4.20	00000	00000	00000	00000	00000	00000	00000	00000
-4.00	00000	00000	00000	00000	00000	00000	00000	00000
-3.80	00000	00000	00000	00000	00000	00000	00000	00000
-3.60	00000	00000	00000	00000	00000	00000	00000	00000
-3.40	00000	00000	00000	00000	00000	00000	00000	00000
-3.20	00000	00000	00000	00000	00000	00000	00000	00000
-3.00	00000	00000	00000	00000	00000	00000	00000	00000
-2.80	00001	00000	00000	00000	00000	00000	00000	00000
-2.60	00000	00000	00000	00000	00000	00000	00000	00000
-2.40	00186	00000	00000	00000	00000	00000	00000	00000
-2.20	01888	00001	00000	00000	00000	00000	00000	00000
-2.00	09255	00009	00000	00000	00000	00000	00000	00000
-1.80	25660	00233	00000	00000	00000	00000	00000	00000
-1.60	47311	02702	00007	00000	00000	00000	00000	00001
-1.40	66191	13908	00198	00008	00003	00003	00007	00026
-1.20	77573	38505	92761	00223	00000	00103	00222	00571
-1.00	81069	68668	16285	03078	01443	01727	02851	05175
-.80	77070	90587	49044	18500	11003	12206	16117	22500
-.60	65141	97233	89816	56423	40530	42330	47579	55535
-.40	47119	88555	1.13369	1.01209	84447	34710	85816	89192
-.20	28610	68233	1.07401	1.20373	1.13796	1.11633	1.06750	1.03598
.00	14786	43834	80625	1.04800	1.10421	1.07504	1.00856	94666
.20	06810	23518	49734	72298	34088	82454	78075	72598
.40	02979	10810	25928	41738	53528	53452	52294	48888
.60	01291	04443	11690	20816	29672	30485	31370	29738
.80	00538	01695	04634	09111	14628	15590	17154	16569
1.00	00196	00594	01621	03519	06440	07158	08572	08458
1.20	00057	00177	00491	01197	02510	02910	03868	03905
1.40	00012	00041	00122	00354	00851	01021	01534	01588
1.60	00002	00007	00023	00088	00244	00300	00514	00549
1.80	00000	00001	00003	00018	00058	00071	00140	00155
2.00	00000	00000	00001	00002	00011	00013	00030	00034
2.20	00000	00000	00000	00000	00001	00002	00004	00006
2.40	00000	00000	00000	00000	00000	00000	00000	00001
2.60	00000	00000	00000	00000	00000	00000	00000	00000
2.80	00000	00000	00000	00000	00000	00000	00000	00000
3.00	00000	00000	00000	00000	00000	00000	00000	00000
3.20	00000	00000	00000	00000	00000	00000	00000	00000
3.40	00000	00000	00000	00000	00000	00000	00000	00000
3.60	00000	00000	00000	00000	00000	00000	00000	00000
3.80	00000	00000	00000	00000	00000	00000	00000	00000
4.00	00000	00000	00000	00000	00000	00000	00000	00000
4.20	00000	00000	00000	00000	00000	00000	00000	00000
4.40	00000	00000	00000	00000	00000	00000	00000	00000
4.60	00000	00000	00000	00000	00000	00000	00000	00000
4.80	00000	00000	00000	00000	00000	00000	00000	00000
5.00	00000	00000	00000	00000	00000	00000	00000	00000

MOSN=		8	SITE #= 1001		SPECTRUM # 1 TO 17				
PROBABILITY OF EXCEEDENCE			PERIOD						
	04000	06500	11000	13000	34000	50000	96000	1.60000	
90000	-1.5863	-1.0764	- .7166	- .5403	- .4698	- .4804	- .5129	- .5624	
50000	- .9525	- .5450	- .2632	- .1110	- .0285	- .0298	- .0396	- .0720	
35000	- .7427	- .3674	- .1099	.0369	.1249	.1278	.1279	.1021	
20000	- .5000	- .1543	.0815	.2244	.3214	.3302	.3441	.3264	
10000	- .2707	.0535	.2758	.4181	.5256	.5411	.5705	.5604	
05000	- .0789	.2292	.4436	.5866	.7040	.7251	.7683	.7643	
03500	.0121	.3119	.5225	.6667	.7884	.8122	.8618	.8604	
02000	.1483	.4351	.6396	.7844	.9125	.9392	.9980	1.0001	
01000	.3116	.5785	.7736	.9199	1.0543	1.0837	1.1517	1.1572	
00500	.4701	.7143	.8984	1.0458	1.1843	1.2152	1.2910	1.2933	
00350	.5489	.7810	.9588	1.1071	1.2476	1.2786	1.3573	1.3669	
00200	.6679	.8816	1.0499	1.1992	1.3414	1.3723	1.4556	1.4670	
00100	.8039	.9974	1.1550	1.3069	1.4507	1.4808	1.5673	1.5806	
00050	.9270	1.1043	1.2533	1.4077	1.5521	1.5808	1.6705	1.6855	
00035	.9854	1.1555	1.3012	1.4573	1.6018	1.6298	1.7203	1.7360	
00020	1.0714	1.2317	1.3725	1.5317	1.6768	1.7032	1.7946	1.8115	
00010	1.1685	1.3194	1.4563	1.6186	1.7642	1.7884	1.8816	1.8995	

6. CASE STUDIES AND AN APPLICATION

6.1 Introduction: Models I, II, III and IV

To study the selected properties of the proposed uniform risk spectra, the spectral amplitudes at six sites were calculated. Fig. 6.1.1 is a map showing the coordinates of the six sites and the two faults. The seismic sources consist of two line faults and of a diffused region. The first line fault is the Sierra Madre-Cucamonga fault of about 100 km in length. The second line fault is a portion of the San Andreas fault from Cajon Pass to the vicinity of San Luis Obispo. The diffused region is a rectangular region running from $117^{\circ} 30' \text{ W}$ to 121° W and from 33° N to $35^{\circ} 30' \text{ N}$. For a detailed description of the seismicity of the faults and of the diffused region, the reader is referred to the data file "SEISMIC.INPUT" presented in Section 5.4. The six sites lie in a North-South direction through the center of the Sierra Madre-Cucamonga fault at distances of roughly 10, 25, 50, 100, 150 and 200 km south of the fault. Table 6.1.1 gives the coordinates of the locations of the six sites.

Four models of seismic risk have been studied. Model I is the original model of seismic risk (Anderson and Trifunac, 1977). It uses the original scaling functions for scaling Fourier and pseudo relative velocity spectra as was done in the original program "EQRISK." The original (old) scaling functions use the Richter's attenuation function. Model II uses the same (old) scaling functions but includes in the calculation the uncertainties in the estimation of seismicity and of maximum allowed earthquake sizes. Model III uses the new frequency dependent attenuation functions which replace the old Richter's attenuation functions in the scaling of Fourier and response spectral amplitudes. Like Model I, Model III does not include uncertainties in estimation of seismicity in the calculations. Finally, Model IV uses new frequency attenuation functions and includes in the

* L.A. VICINITY SITES # 1. 2. 3. 4. 5. 6

FAULT #1 SIERRA MADRE - CUCAMONGA FAULT ZONE

FAULT #2 SAN ANDREAS FAULT - CAJON PASS TO SAN LUIS OBISPO

DIFFUSE REGION:

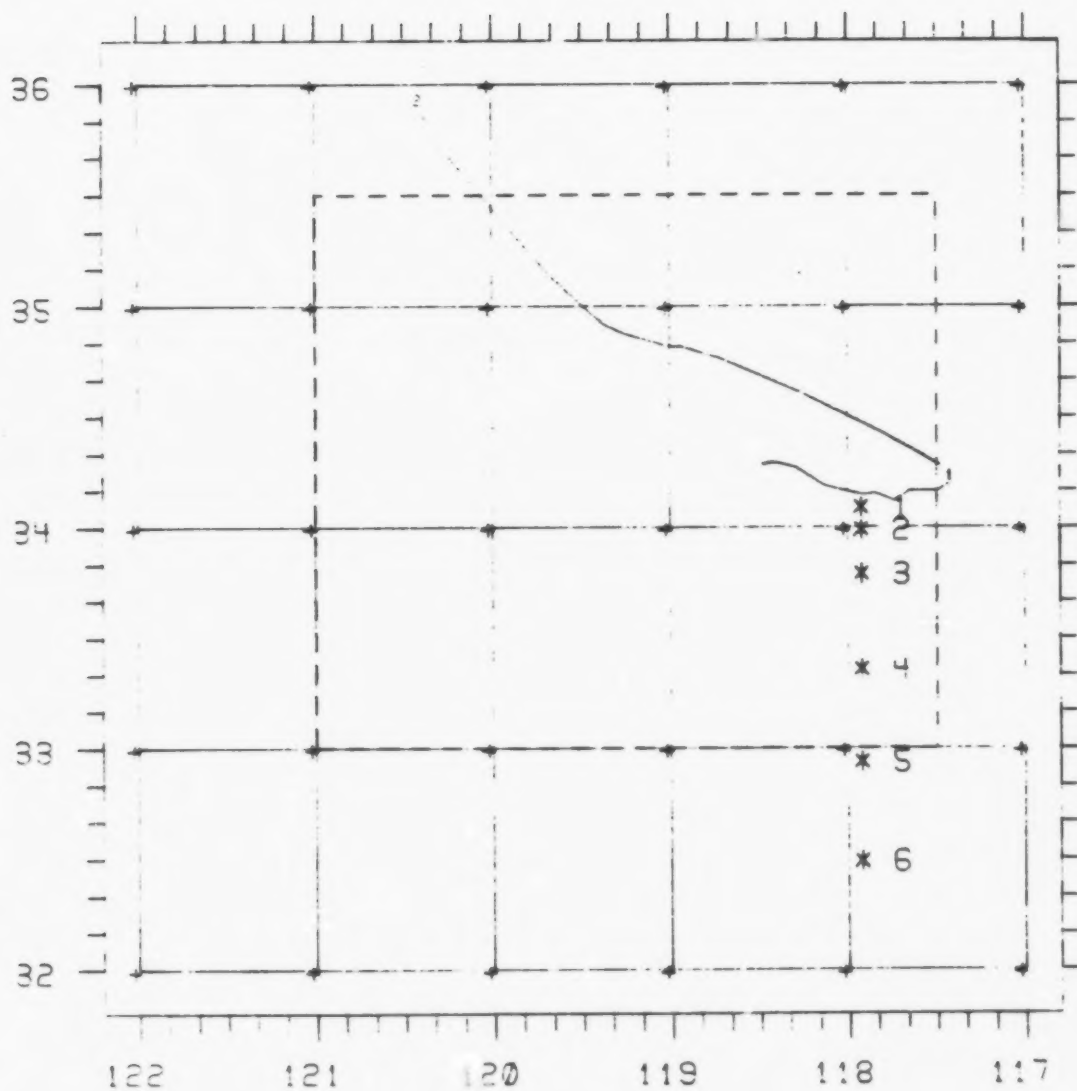


Table 6.1.1

SITE #	LATITUDE (N)	LONGITUDE (W)
1	34.1°	117.9°
2	34.0°	117.9°
3	33.8°	117.9°
4	33.367°	117.9°
5	32.95°	117.9°
6	32.50°	117.9°

Table 6.1.2

Model	Scaling Functions	Includes Uncertainties Seismicity
I	OLD	NO
II	OLD	YES
III	NEW	NO
IV	NEW	YES

calculations the uncertainties in the estimation of seismicity and of maximum allowed earthquake sizes. Table 6.1.2 is a summary of these characteristics of the four models.

6.2 Seismic Risk Spectra at 6 Sites

Figures 6.2.1 through Figure 6.2.4 show the seismic risk spectra of the four models at the six sites. Each figure gives the expected pseudo relative velocity amplitudes $PSV(T)$, at 8 periods T between .04 sec. and 2 sec., that will be exceeded at the 6 sites for the given probability levels of $P = 0.1, 0.5$ and 0.9 . Each curve in the figure is associated with the corresponding site number (1 to 6) it represents. Thus the top six solid curves represent the PSV amplitudes that will be exceeded at the six sites with the probability level of $P = 0.1$. The next six dashed curves in the middle correspond to the corresponding amplitudes at $P = 0.5$. The bottom six solid curves correspond to $P = 0.9$.

Sites #1 through #6 are at 10, 25, 50, 100, 150 and 200 km from the nearest fault, the Sierra Madre-Cucamonga fault. All four models show that, at the same probability level, the expected PSV amplitudes are of decreasing amplitudes from site #1 down to site #6. The PSV amplitudes in all four figures are presented to periods of only up to 2 seconds, because for periods longer than 2 seconds, the PSV amplitudes may be distorted by noise, especially at low probability levels of exceedance.

6.3 Comparison of Different Models

It is interesting to compare the results of the different models at various sites. Figure 6.3.1 shows a comparison of the PSV spectra at site #1 for 5 probability levels $P = 0.1, 0.2, 0.35, 0.5$ and 0.9 . The five dashed lines correspond to the PSV amplitudes (for $p = 0.1$ to 0.9)

PSV SEISMIC RISK SPECTRA AT 6 SITES
FOR $P = 0.1, 0.5 \text{ \& } 0.9$
MODEL I: ORIGINAL SCALING FUNCTIONS

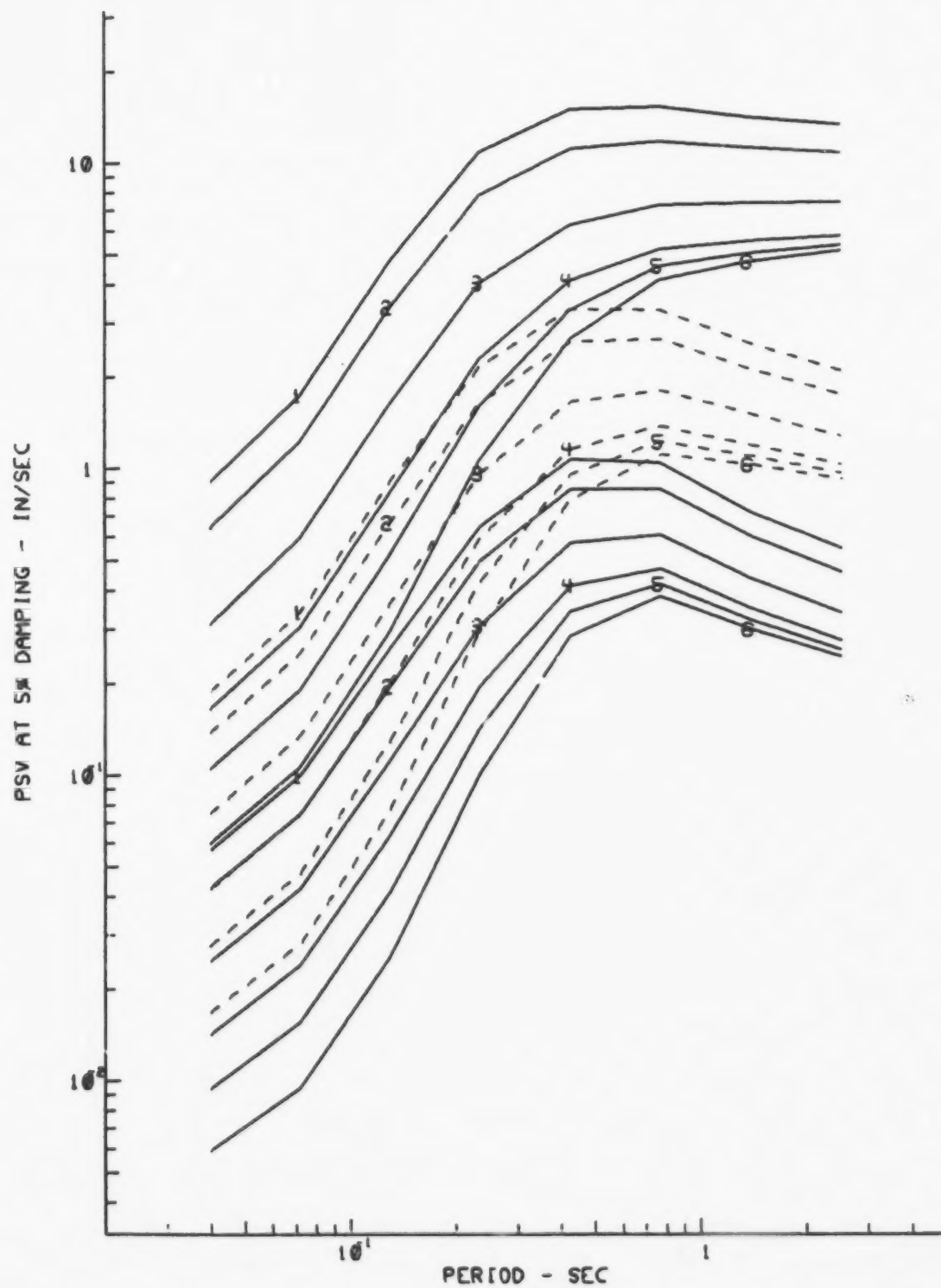
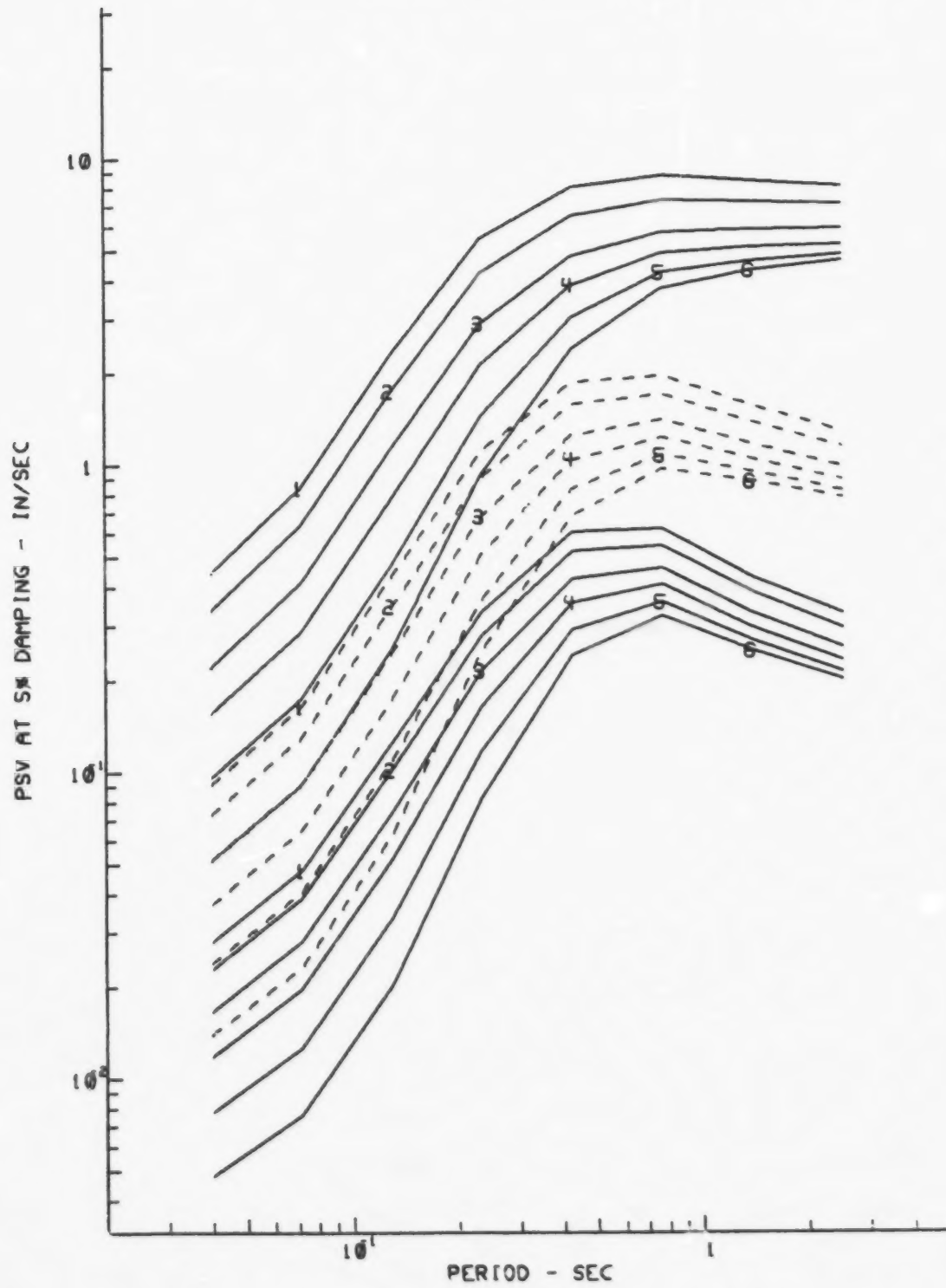


Figure 6.2.1

BEST COPY AVAILABLE

PSV SEISMIC RISK SPECTRA AT 6 SITES

FOR $P = 0.1, 0.5 \text{ \& } 0.9$
 MODEL II: ORIGINAL SCALING FUNCTIONS
 UNCERTAINTIES IN SEISMICITY INCLUDED



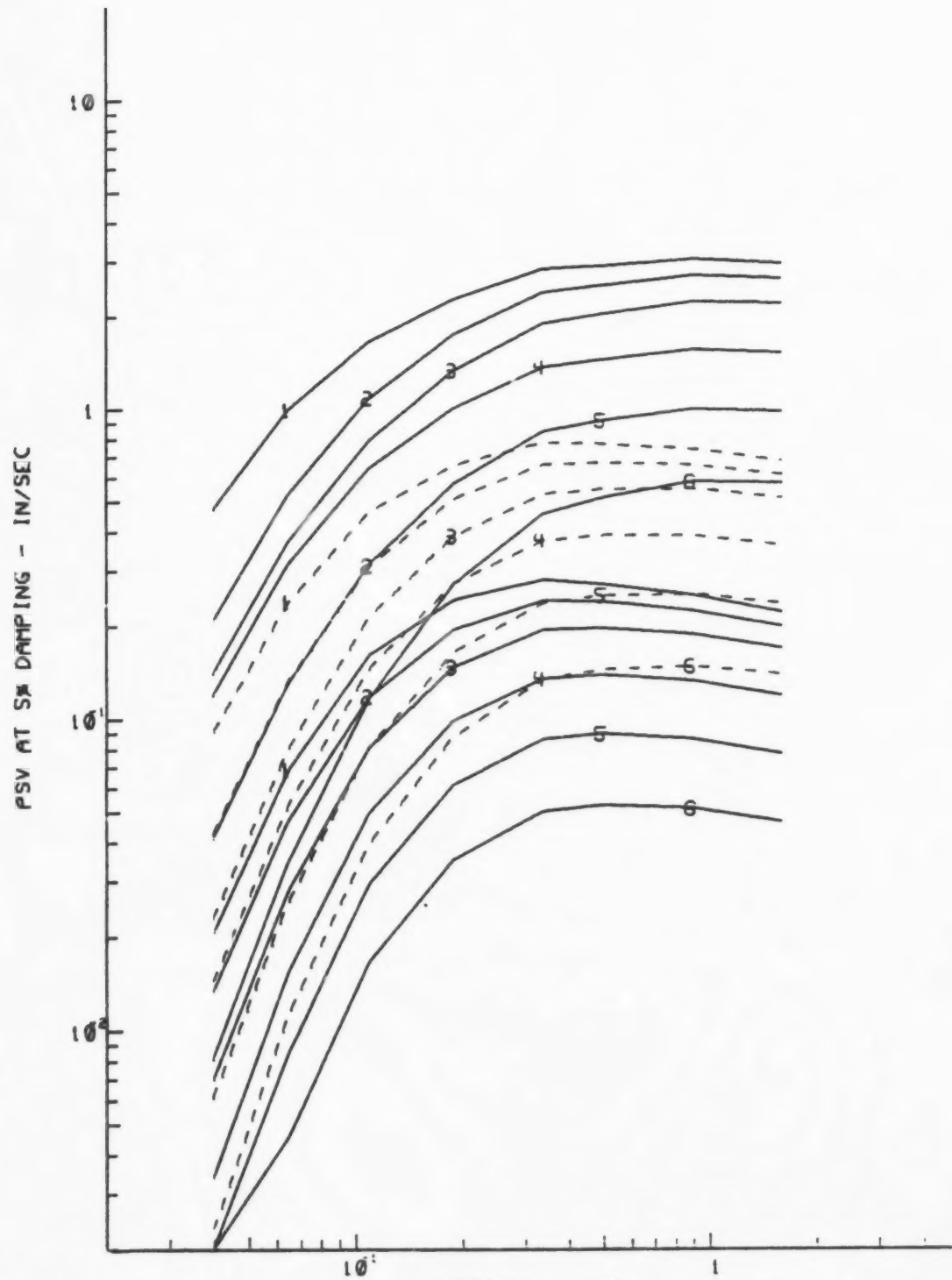
PERIOD - SEC

Figure 6.2.2

BEST COPY AVAILABLE

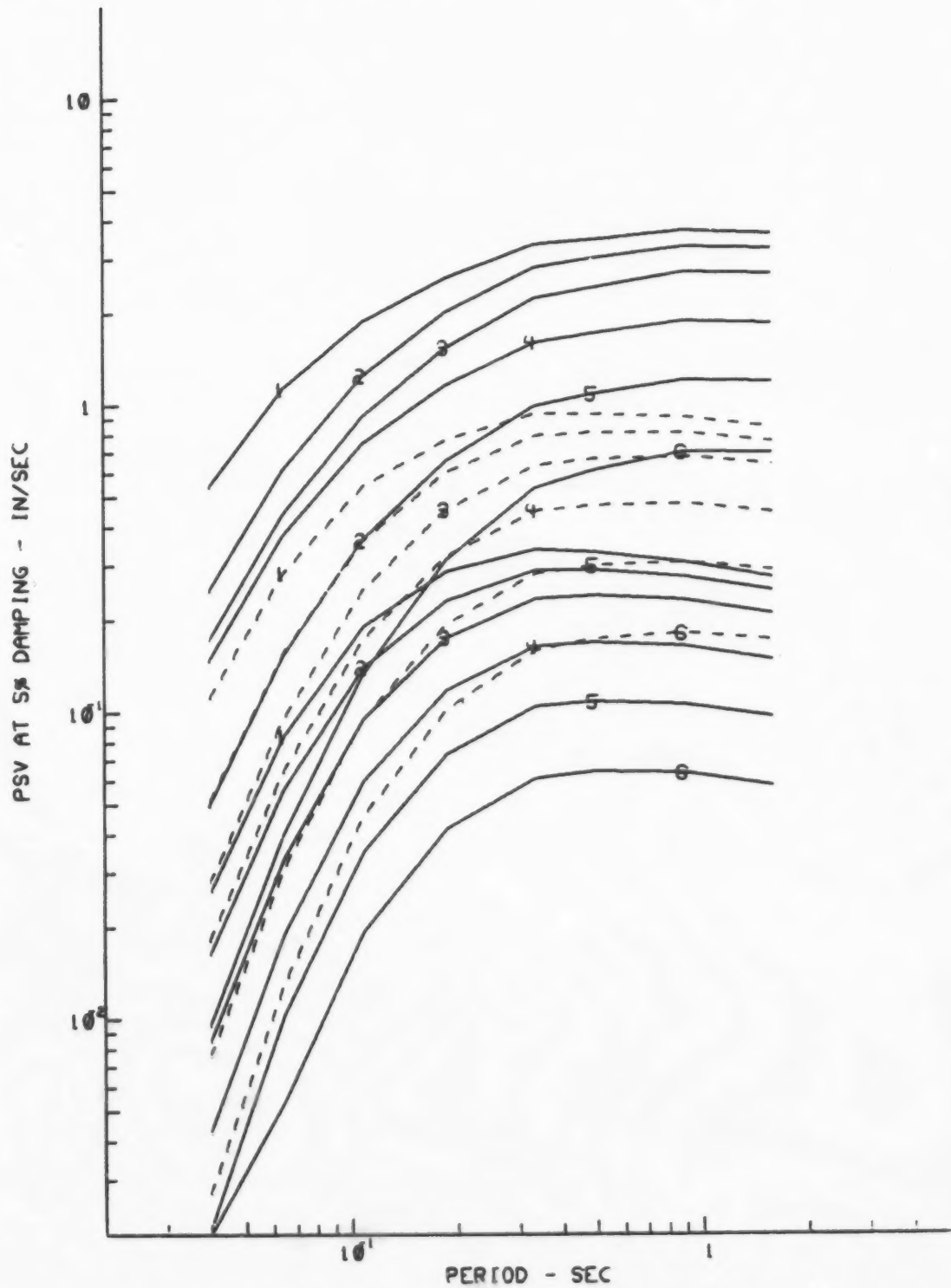
PSV SEISMIC RISK SPECTRA AT 6 SITES

FOR $P = 0.1, 0.5 \text{ \& } 0.9$
MODEL III: NEW SCALING FUNCTIONS



PSV SEISMIC RISK SPECTRA AT 6 SITES

FOR $P = 0.1, 0.5 \text{ \& } 0.9$
 MODEL IV: NEW SCALING FUNCTIONS
 UNCERTAINTIES IN SEISMICITY INCLUDED



PERIOD - SEC

Figure 6.2.4

BEST COPY AVAILABLE

PSV SEISMIC RISK SPECTRA AT SITE #1

FOR $P = 0.1, 0.2, 0.35, 0.5 \text{ \& } 0.9$

ORIGINAL SCALING FUNCTIONS

UNCERTAINTIES IN SEISMICITY

MODEL I (-----): NOT INCLUDED

MODEL II (SOLID): INCLUDED

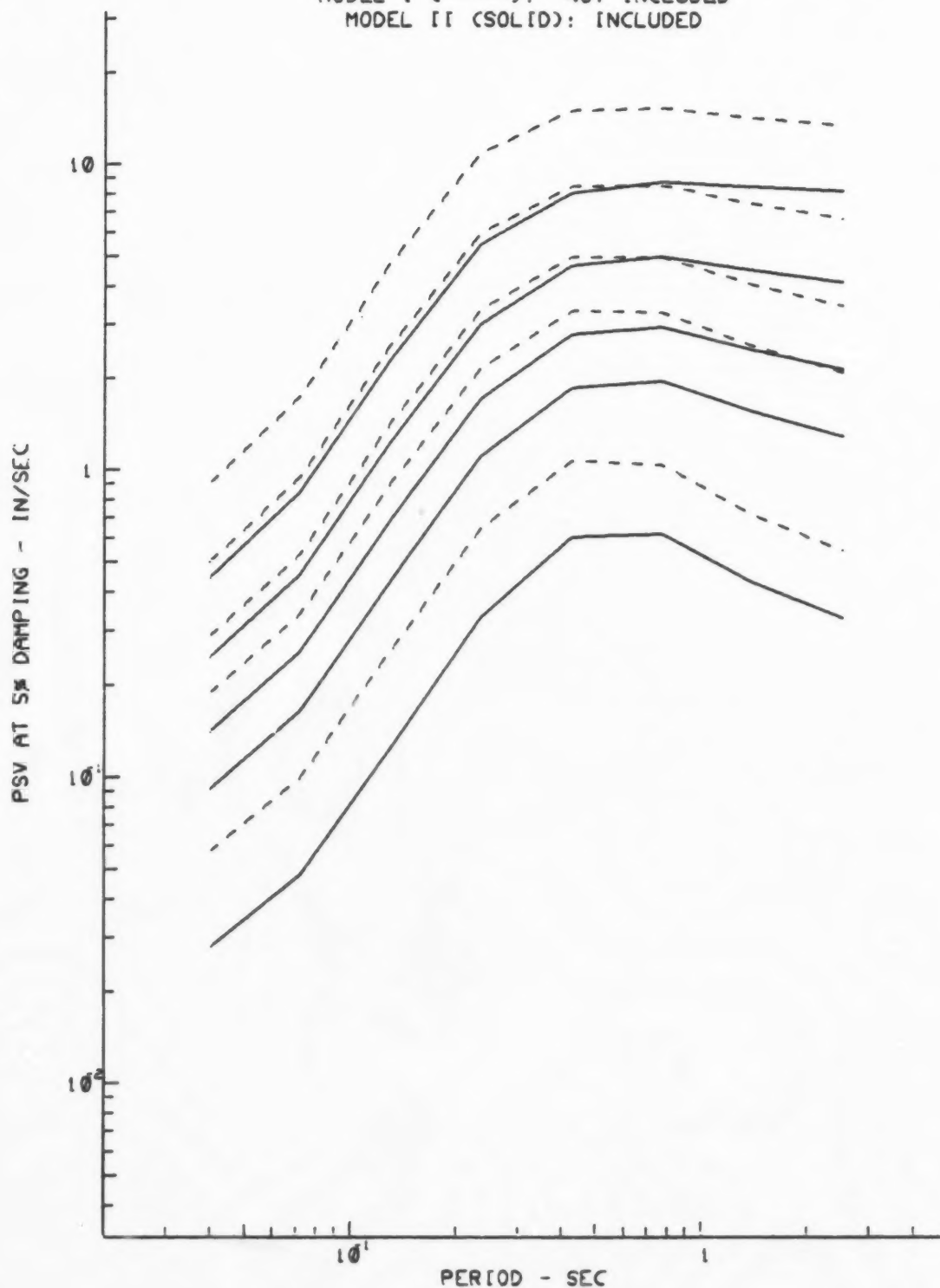


Figure 6.3.1

PSV SEISMIC RISK SPECTRA AT SITE #6

FOR $P = 0.1, 0.2, 0.35, 0.5 \text{ \& } 0.9$

UNCERTAINTIES IN SEISMICITY:

MODEL I (-----): NOT INCLUDED

MODEL II (SOLID): INCLUDED

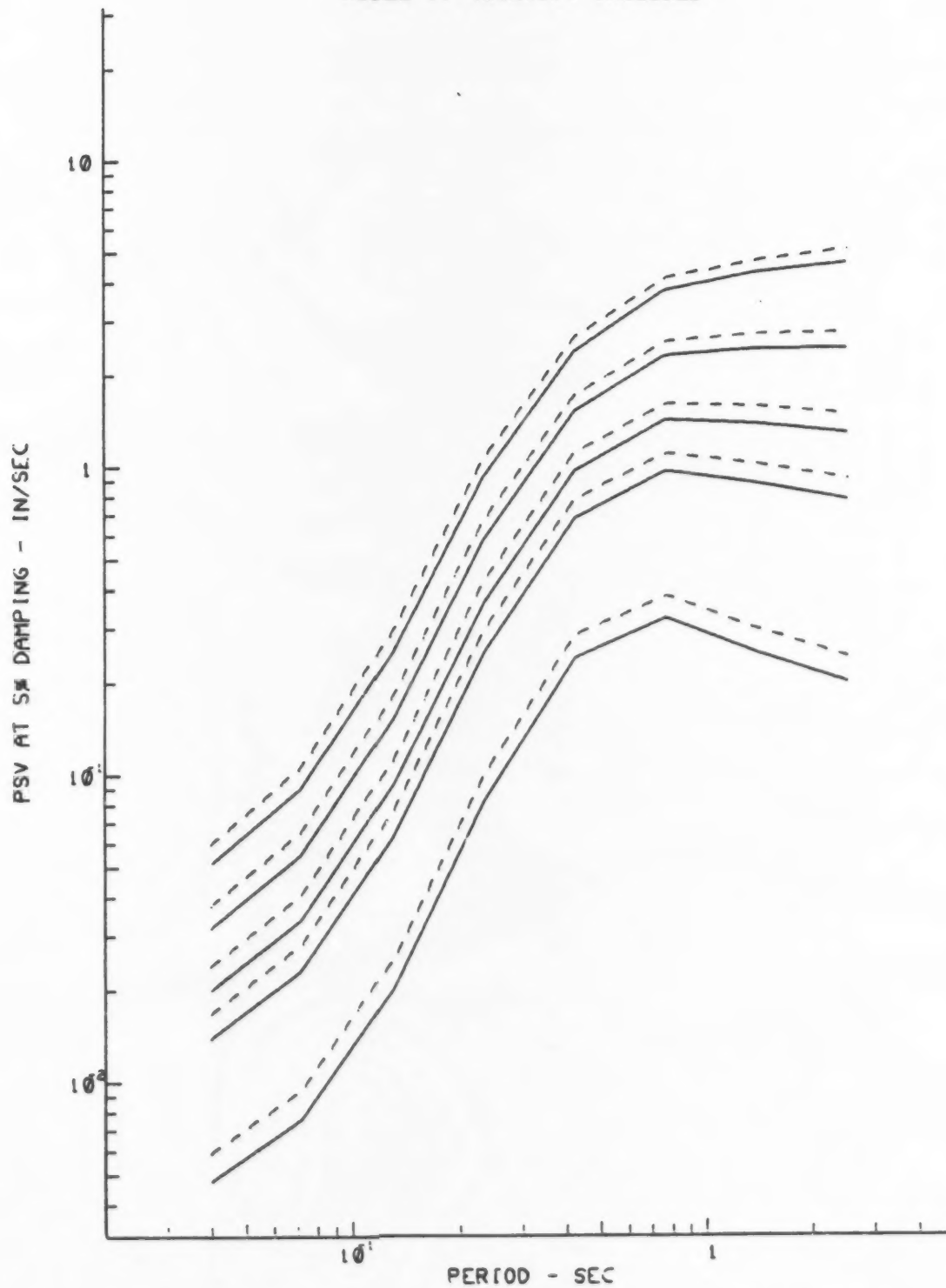


Figure 6.3.2

BEST COPY AVAILABLE

PSV SEISMIC RISK SPECTRA AT SITE #1

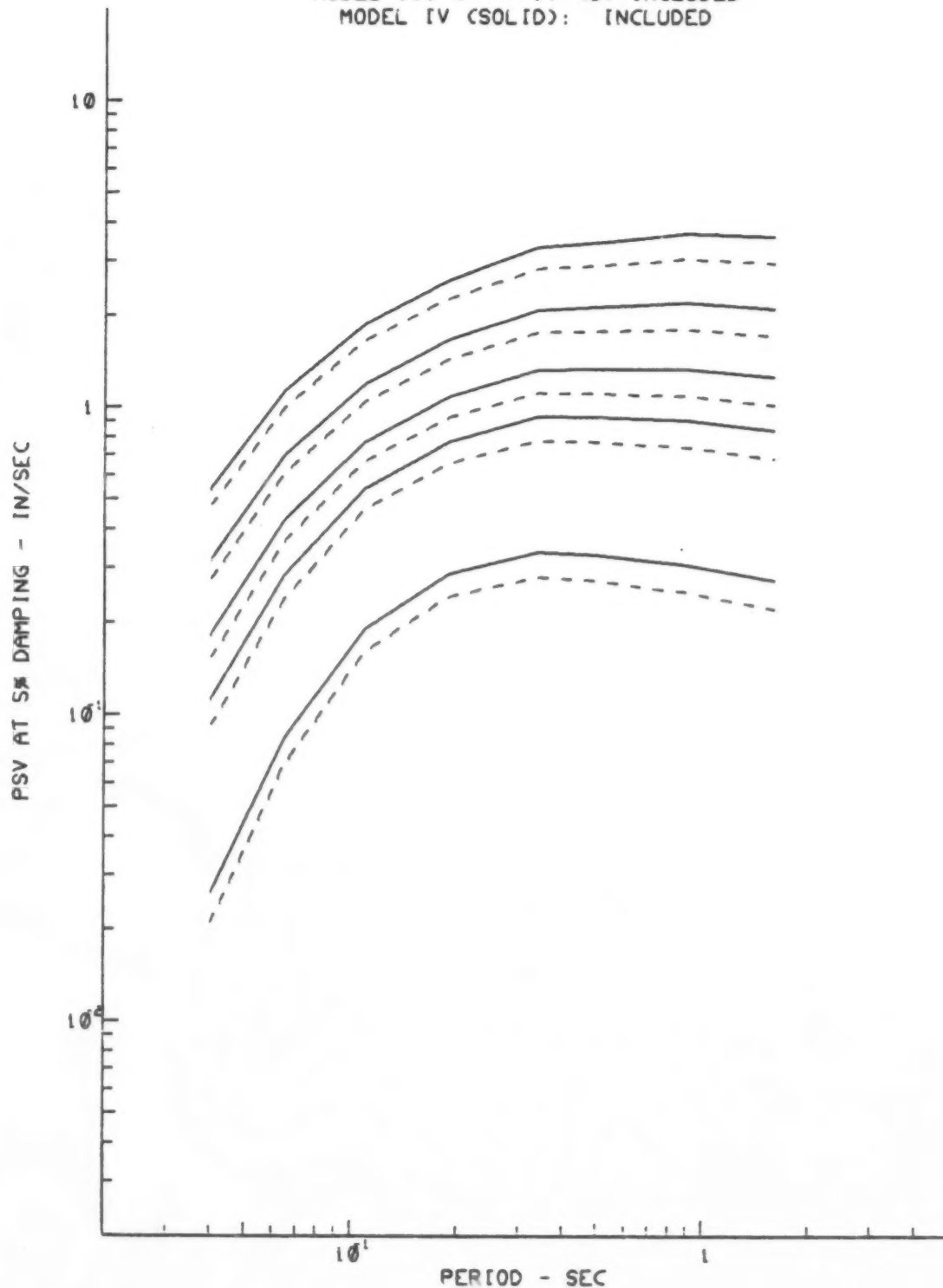
FOR $P = 0.1, 0.2, 0.35, 0.5 \text{ \& } 0.9$

NEW SCALING FUNCTIONS

UNCERTAINTIES IN SEISMICITY

MODEL III (-----): NOT INCLUDED

MODEL IV (SOLID): INCLUDED



BEST COPY AVAILABLE

Figure 6.3.3

169

PSV SEISMIC RISK SPECTRA AT SITE #6

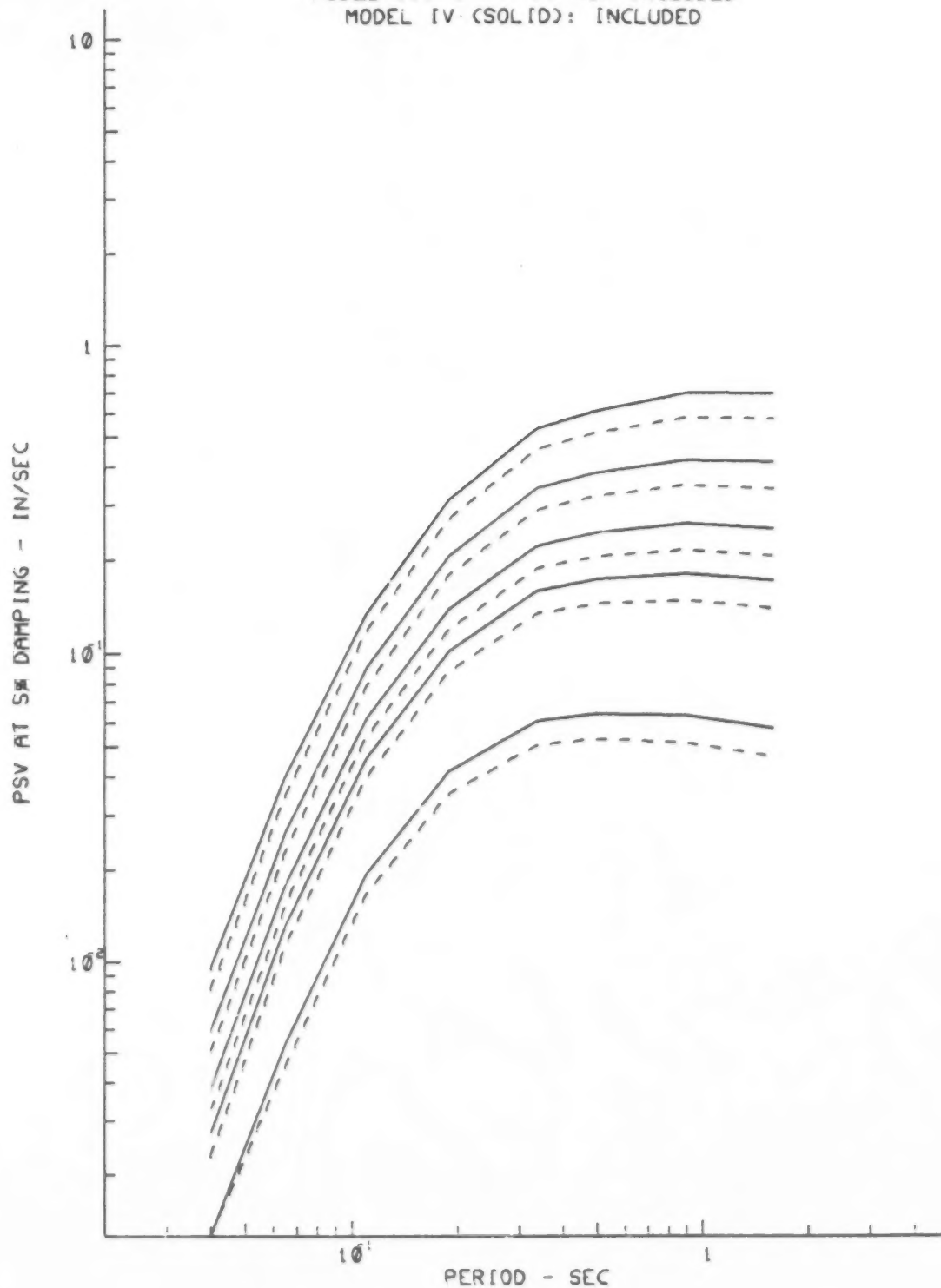
FOR $P = 0.1, 0.2, 0.35, 0.5 \text{ \& } 0.9$

NEW SCALING FUNCTIONS

UNCERTAINTIES IN SEISMICITY:

MODEL III (---): NOT INCLUDED

MODEL IV (SOLID): INCLUDED



PERIOD - SEC

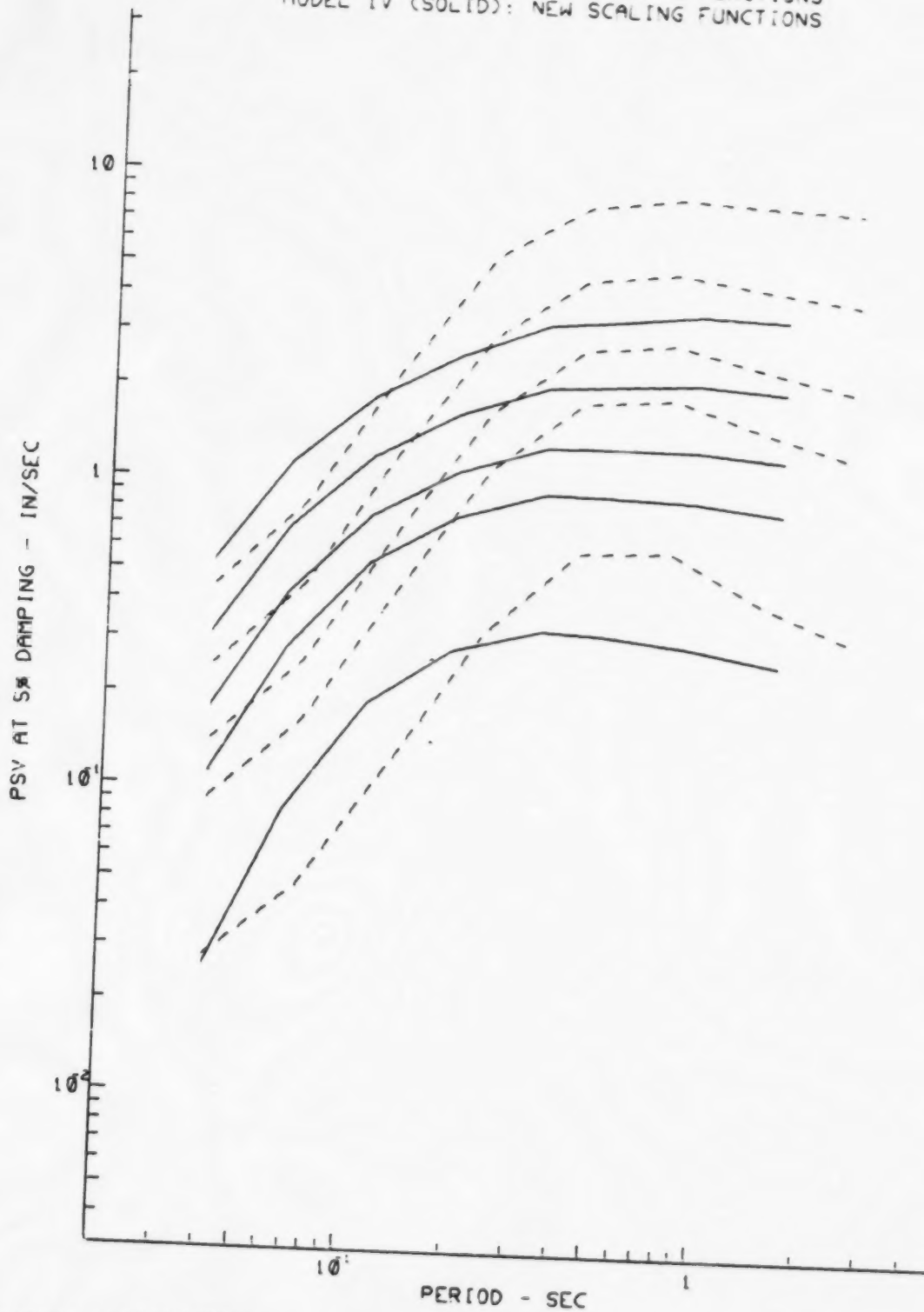
Figure 6.3.4

170

BEST COPY AVAILABLE

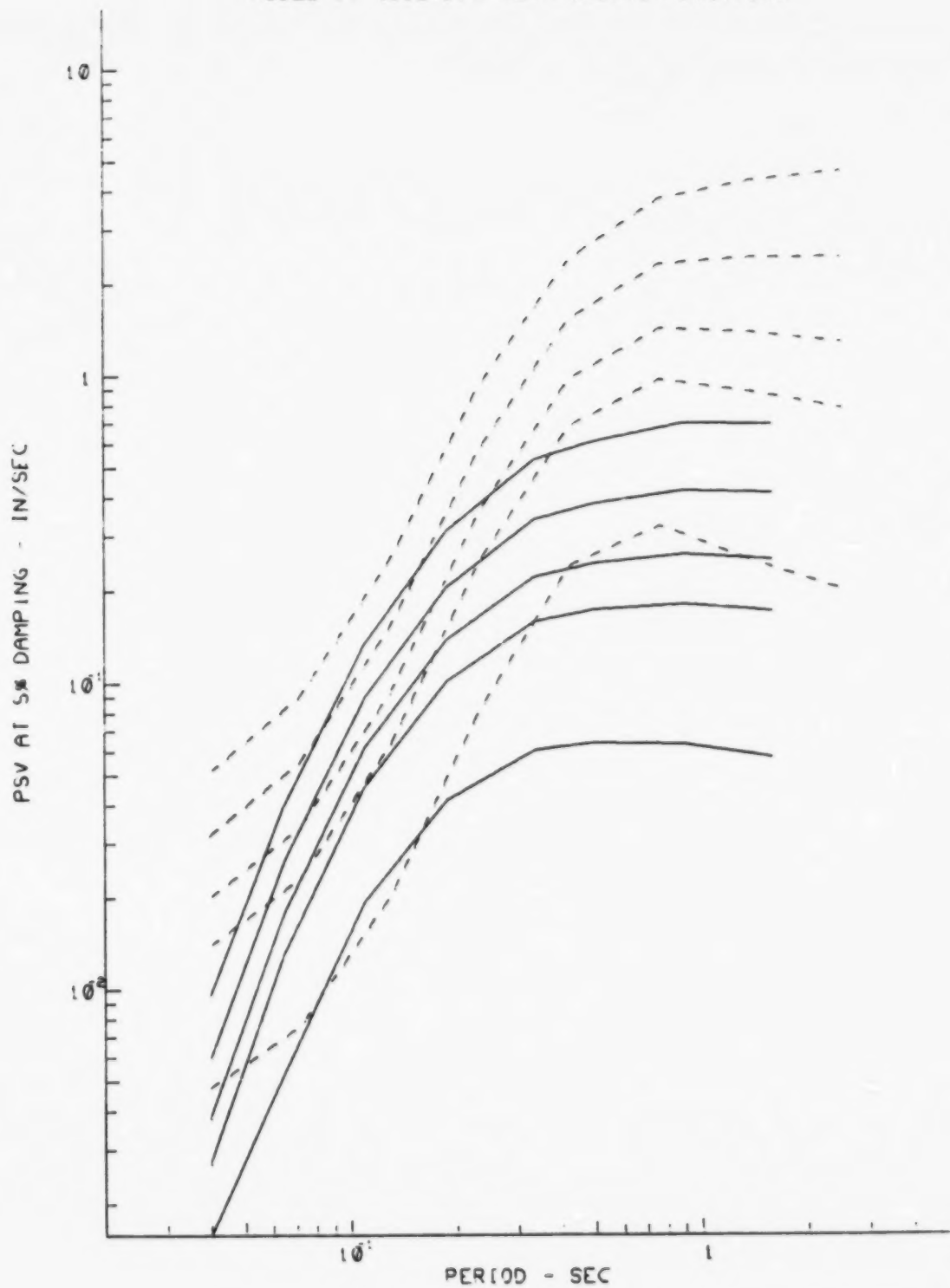
PSV SEISMIC RISK SPECTRA AT SITE #1

FOR $P = 0.1, 0.2, 0.35, 0.5 \text{ \& } 0.9$
UNCERTAINTIES IN SEISMICITY INCLUDED
MODEL II (-----): OLD SCALING FUNCTIONS
MODEL IV (SOLID): NEW SCALING FUNCTIONS



PSV SEISMIC RISK SPECTRA AT SITE #6

FOR $P = 0.1, 0.2, 0.35, 0.5 \text{ \& } 0.9$
UNCERTAINTIES IN SEISMICITY INCLUDED
MODEL II (-----): OLD SCALING FUNCTIONS
MODEL IV (SOLID): NEW SCALING FUNCTIONS



PERIOD - SEC

Figure 6.3.6

at the five probability levels for Model I, while the five solid lines correspond to those for Model II. Both Models I and II use the old scaling functions with Richter's attenuation function and the old database. The Model II differs from Model I in that Model II includes the uncertainties in the estimation of seismicity and of maximum allowed earthquake size in the calculation of the risk spectra. Figure 6.3.1 shows that for each of the five probability levels shown, the estimated PSV levels from Model I are higher than those from Model II in the entire period range. Figure 6.3.2 shows the same comparison of the two models but at site #6. The PSV spectra from Model I again are higher than the corresponding spectra from Model II, but this time to a smaller degree.

These two figures might lead one to conclude that, by including the uncertainties in the estimation of seismicity and in maximum allowed earthquake size into the calculation of risk spectra, the resulting amplitudes (Model II) at each level of probability may actually be lower than the corresponding spectra (Model I) computed without consideration of such uncertainties. Also the two figures might suggest that the differences may be larger for sites closer to the fault (site #1) than those further away (site #6).

6.4 An Application

The following example is presented to illustrate the properties of the proposed model in a realistic practical setting. As in Anderson and Trifunac (1977), an application is made to evaluate the seismic risk for a site on the north coast of Puerto Rico, a site which has been studied by Kelleher et al. (1973), and by Der-Kiureghian and Ang (1975). To allow for comparison of the present model with that presented in Anderson and Trifunac (1977), the seismicity of the area will be described using the data available then

to Der-Kiureghian and Ang (1975), rather than obtaining a more current listing of events and their epicenters. The data used consists of a list of all events in the magnitude range from 3.0 to 8.5 that are known to have occurred in the area between 1915 and 1971. The area is within the rectangle bounded by latitudes 17°N to 21°N and longitudes 63°W to 70°W .

On the basis of the available events and using additional information in Kelleher et al. (1973), four models of the region have been constructed, each describing the seismicity to be expected in the region in the next 50 years. These four models, labeled A, B, C and D, are reconstructed here following Trifunac and Anderson (1977), where the models have been described independently by four experts, each describing the seismicity they expected in the same region during the next 50 years. The models are shown in Figure 6.4.1 and are identical to those in Figure 2.1.4 of Anderson and Trifunac (1977). The following is a brief summary of how the four models have been constructed:

Model A assumes that the seismicity during the next 50 years will essentially be the same as that which has occurred in the past. The map is divided into five zones: I, II, III, IV and V. Zones I through IV are subregions while zone V is the whole region (17° to 21°N , 63° to 70°W) overlapping with the four subregions. The overall seismicity is first assigned to the whole region (zone V), and additional seismicity is then assigned to the four subregions (zones I, II, III and IV). The seismicity in each zone is scaled to account for the differences in the time periods for which the data is available and the period for which the risk is to be estimated (50 years). This model will not consider the incomplete coverage of small earthquakes for much of the historical record.

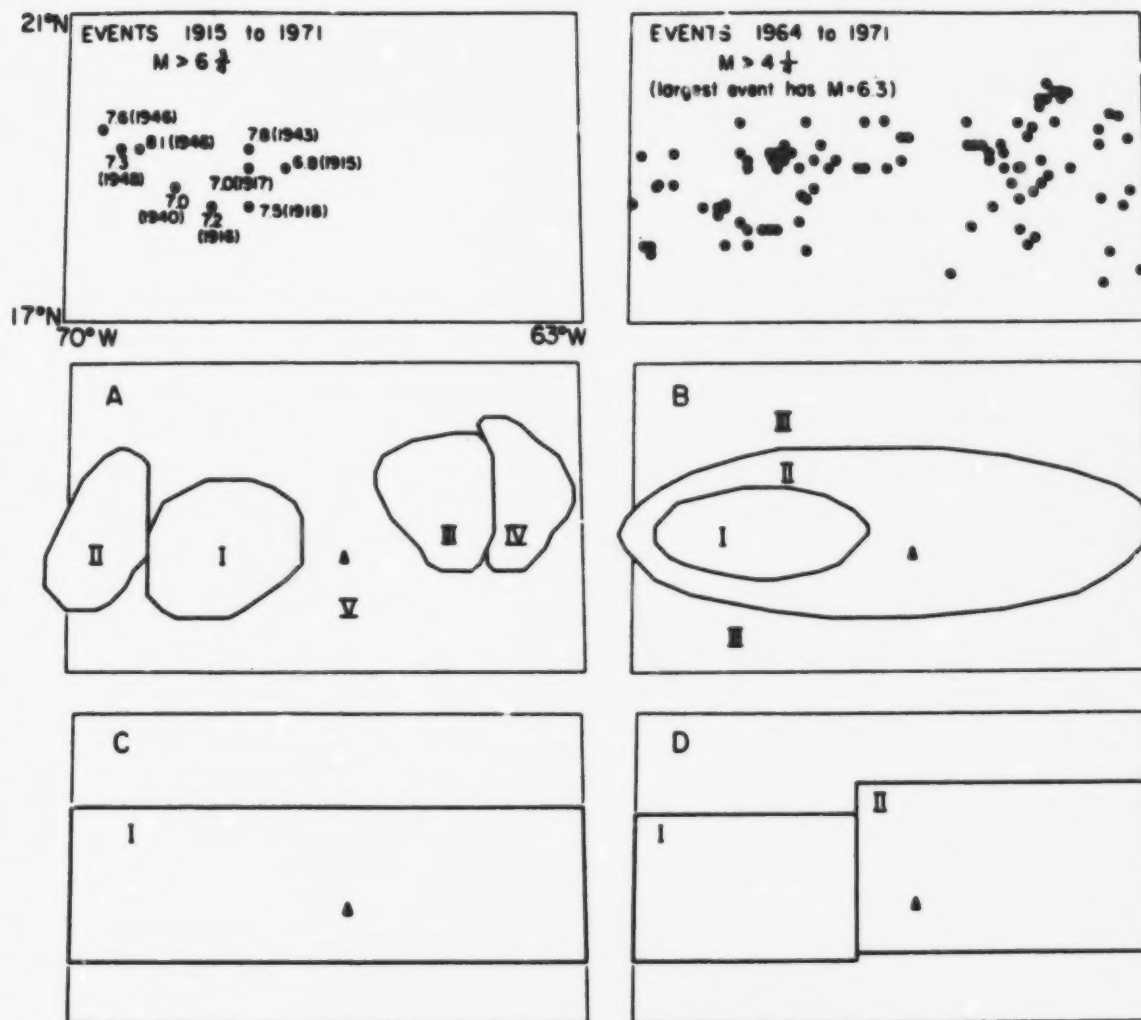


Figure 6.4.1 Seismicity in the vicinity of Puerto Rico, as given by Der-Kiureghian and Ang (1975), and 4 models to describe this seismicity. The boundary of all 6 maps is the same as shown in the upper left. The map at the upper left shows the epicenters of major earthquakes since 1915; the map at the upper right shows events with magnitude greater than $4 \frac{1}{4}$ which occurred between 1964 and 1971. The estimated seismicity rates in each zone of the 4 models (A, B, C, D) are in Table 6.4-1. These zones are the projections of the dipping planar source region for the risk estimates. Models are for the site shown by a solid triangle.

Model B assumes that the major seismic activity in the next 50 years will continue to occur where the major activity of the past was recorded. The map is divided into three zones: I, II and III. Zones I and II are subregions while zone III is again the whole region. As in Model A, the overall seismicity is first assigned to the whole region (zone III), and additional seismicity is then assigned to the two subregions (zones I and II). This model emphasizes the data on smaller events as recorded in recent years.

Model C assumes that the entire seismic activity is associated with an active plate margin, and makes no differentiation between the western portion which has a history of large events and the eastern portion which does not have that history. There is thus only one uniform zone for the whole region.

Model D consists of two zones: zone I for the western region and zone II for the eastern region. It assumes that because, as Kelleher et al. (1975) point out, the eastern region (zone IV) has not had any large historic earthquakes, that this portion of the plate margin must be considered a more likely candidate for a large earthquake in the future. The western region (zone I) was assigned the seismicity which is lower than the historic rates on the assumption that some of the strains there were already relieved by the past events, and would take some time to accumulate to a triggering level again.

The estimated seismicities in each zone of the four models (A,B,C and D) are given in Table 6.4-1. These models will be used to calculate the uniform risk spectra at a site shown in Figure 6.4.1.

TABLE 6.4-1

Number of Earthquakes assigned to the source zones of Figure 2.14
for the Puerto Rico applications.

Case	Zone	Area 10^4 km^2	Magnitude											
			3.0	3.5	4.0	4.5	5.0	5.5	6.0	6.5	7.0	7.5	8.0	8.5
A	I	3.83	91	59	38	25	16	10	7	4	3	2	1	0
	II	2.80	23	17	12	9	6	5	3.3	2.4	1.7	1.3	0.9	0
	III	2.56	58	35	22	13	8	5	3	2	0	0	0	0
	IV	1.78	83	44	23	12	6	3	2	1	0	0	0	0
	V	32.61	107	54	28	14	7	4	2	1	0	0	0	0
B	I	3.02	3583	1095	319	146	44	29	13	7	5	3	0.5	0
	II	14.44	13555	4142	1209	554	166	111	50	0	0	0	0	0
	III	32.62	20779	6350	1852	0	0	0	0	0	0	0	0	0
C	I	16.33				66	40	23	13	6	5	3	2	1
D	I	6.83				501	178	63	22	8	3	1	0	0
	II	10.26				889	315	112	40	14	5	3	1.5	0

In calculating the uniform risk spectra for the site, variations of each of these models have been considered:

Variation 1 assigns each of the zones of each model to be a diffuse region on the surface. To allow for comparison of the present model with the previous ones, the length of rupture is not included here as in Anderson and Trifunac (1977). This means that each event will be assumed to have rupture length zero, or equivalent, each event is assumed to be a point source, with equal probability of occurring anywhere within the specified zone.

Variation 2 assigns each of the zones of each model to be a diffuse region on a plane dipping south from the northern edge of the region at an angle of 35 degrees. This variation is probably more realistic in describing the geometry of the seismic zone. In the old model of "EQRISK," as was pointed out in Anderson and Trifunac (1977), an approximate extension of the model for attenuation was made for this case. This was done by replacing the epicentral distance with the hypocentral distance. It was emphasized then that this is an approximation. In the present new model of "NEQRISK," as we noted in Section 3.2, a frequency dependent attenuation function, $\Delta t t(\Delta, M, T)$ is employed and a new function $\Delta = \Delta(S, H, R)$, (equation 3.2.2), known as the "representative distance" from the earthquake source of size S , at depth H and epicentral distance R from the recording site is used.

Figures 6.4.2 and 6.4.3 show uniform risk spectra for the seismicity models for variations 1 and 2, calculated using the old "EQRISK" program. Figures 6.4.4 and 6.4.5 present the uniform risk spectra for the same seismicity models but for variations 1 and 2 calculated using the new "NEQRISK" program, with the new frequency dependent attenuation function $\Delta t t(\Delta, M, T)$ and the new "representative" source-to-station distance Δ .

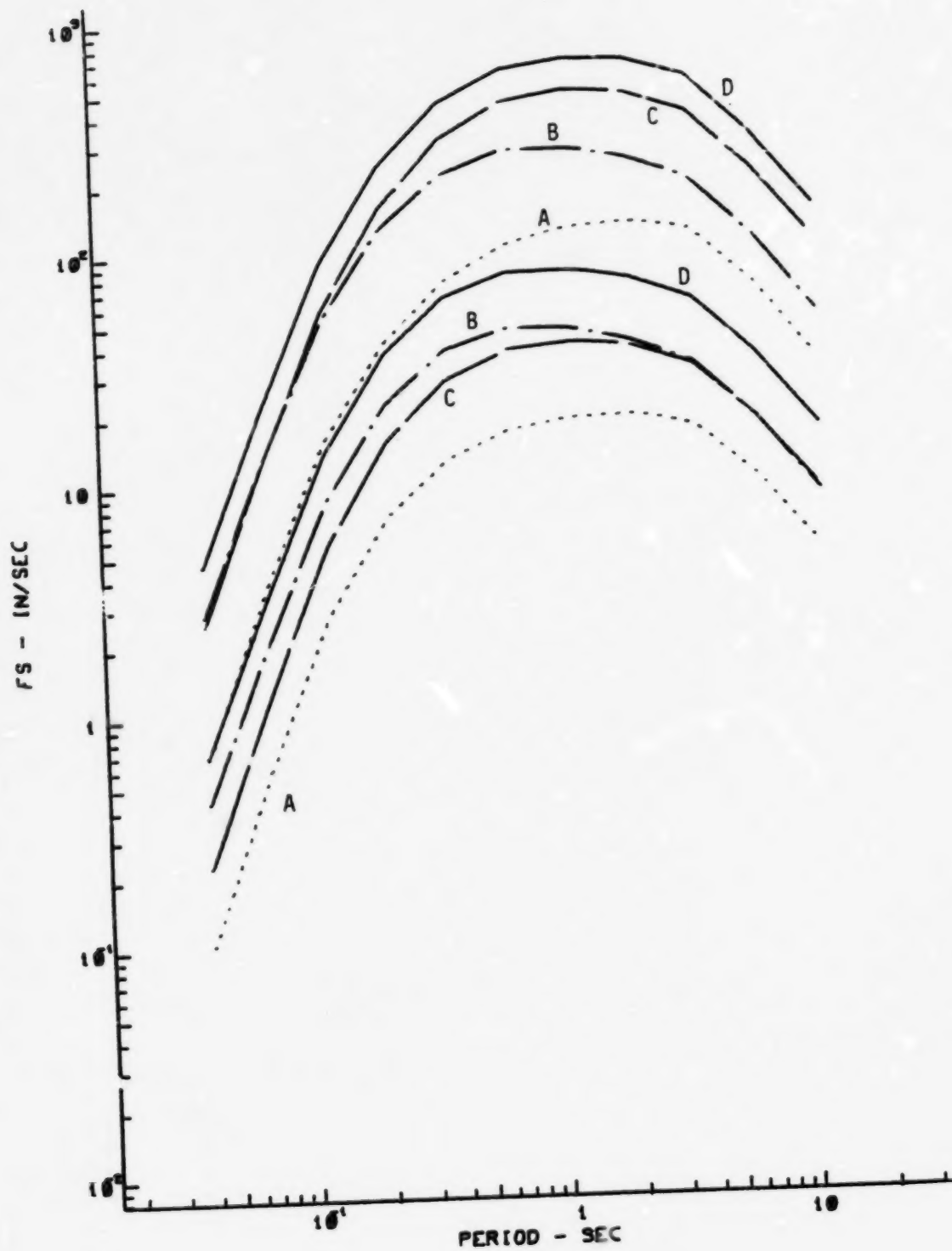


Figure 6.4.2

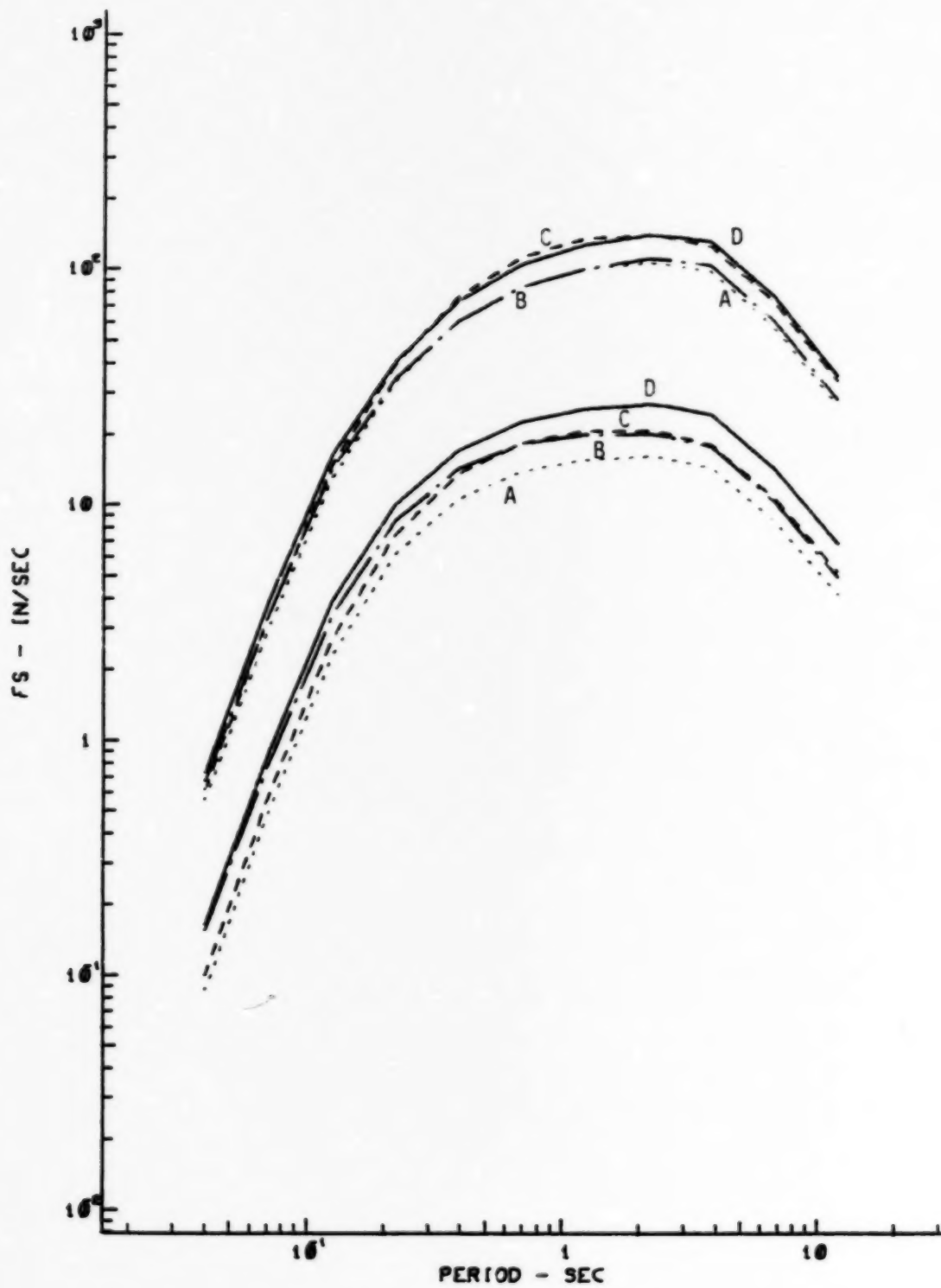


Figure 6.4.3

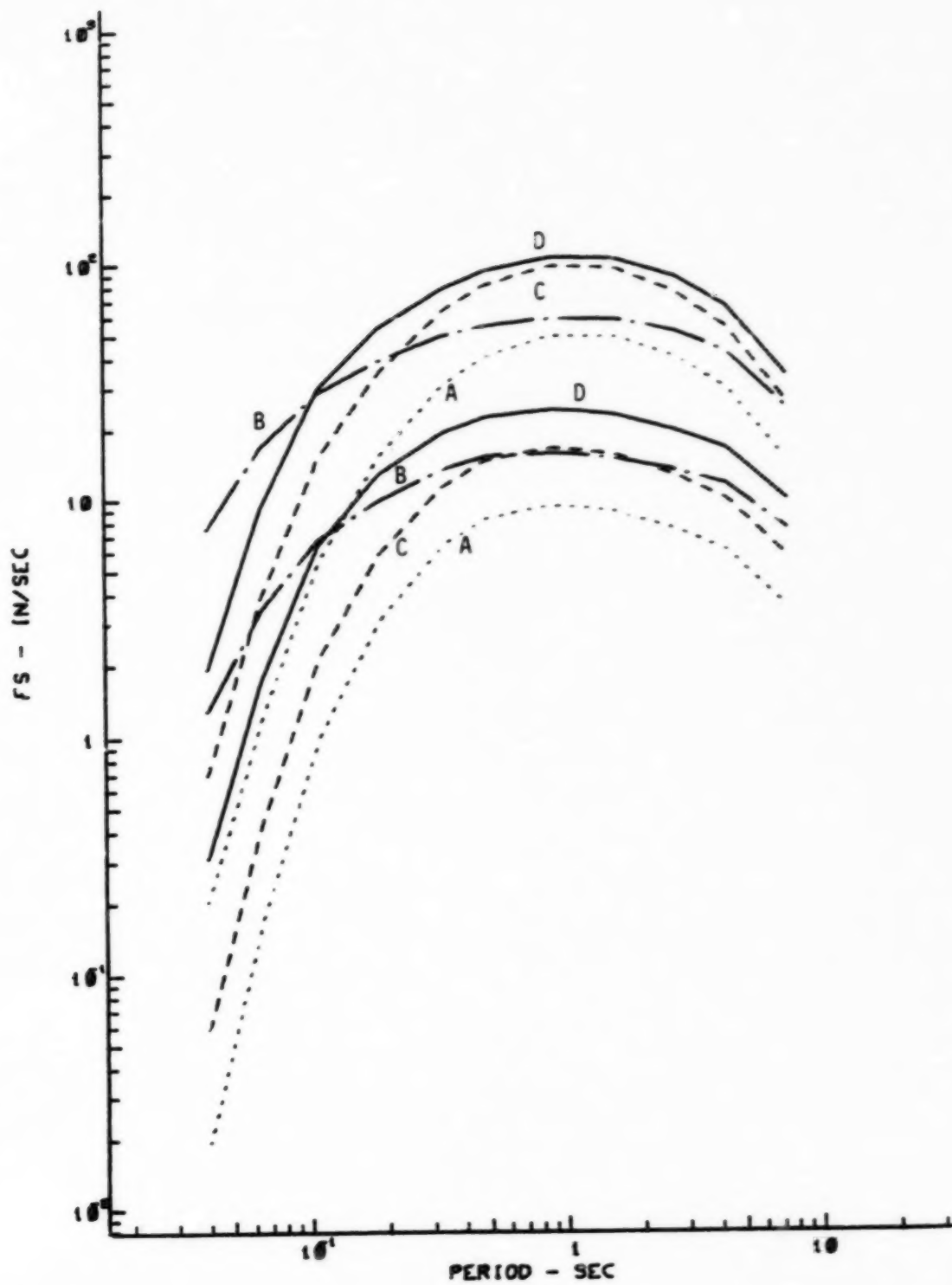


Figure 6.4.4

181

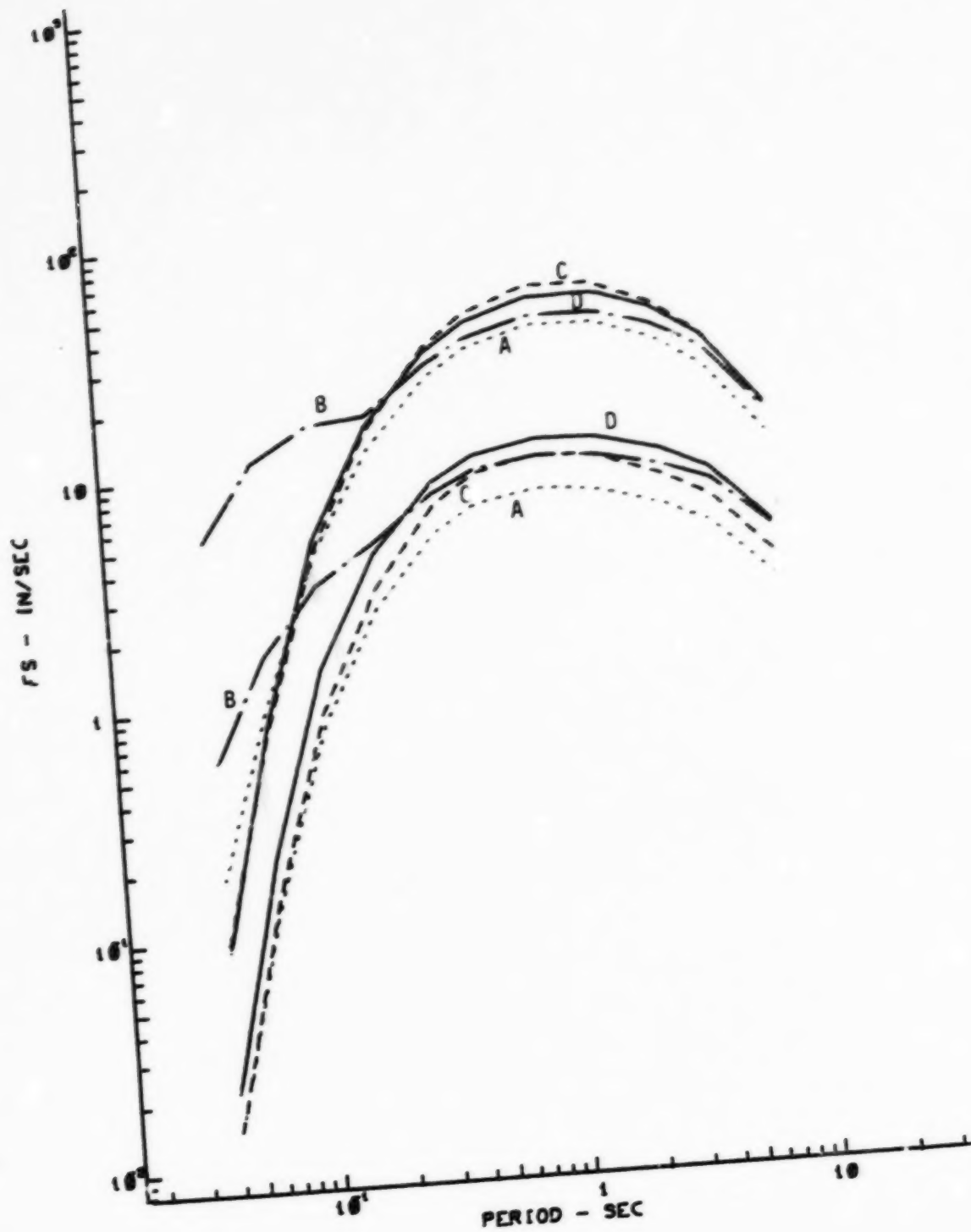


Figure 6.4.5

BEST COPY AVAILABLE

Several observations can be made about the spectra in the four figures 6.4.2 through 6.4.5. First, in the spectra for the old scaling with the "EQRISK" program (Figures 6.4.2 and 6.4.3), the differences between the spectral amplitudes resulting from cases A-D for a given probability level are considerably greater for the diffuse source, Variation 1, than for the dipping plane source, Variation 2. This is due to the fact that the large events, which contribute most to the spectral amplitudes, are in a wider distance range when they are placed on the surface (Variation 1) than when they are placed on the dipping plane (Variation 2). Similar observations can be made about the spectra in Figure 6.4.4 and 6.4.5, for the same seismicity models, for Variations 1 and 2, but calculated by the new "NEQRISK" program.

In using the old attenuation function, it is noted that Model A is depleted in spectral amplitudes of intermediate and high frequencies relative to the other three models (Figures 6.4.2 and 6.4.3), in particular for Variation 1. This is because Model A assumes a small number of small events close to the site. In using the new frequency dependent attenuation function, the similar observation can be made about Model A (Figures 6.4.4 and 6.4.5). It is also noted that Model B has large amplitudes in the high frequency range beyond 10 Hz, when compared to the amplitudes of Models A, C and D. Model B is unique, compared with other models, in that the expected number of small events in the region close to the site is substantially higher for this than for all other models (Table 6.4-1). The new attenuation function $\alpha(\Delta, M, T)$ attenuates faster than the Richter's attenuation function in the distance range below 50 to 100 km and for high frequencies. This results in Model B having risk spectra with higher amplitudes in that period range and from small nearby earthquakes.

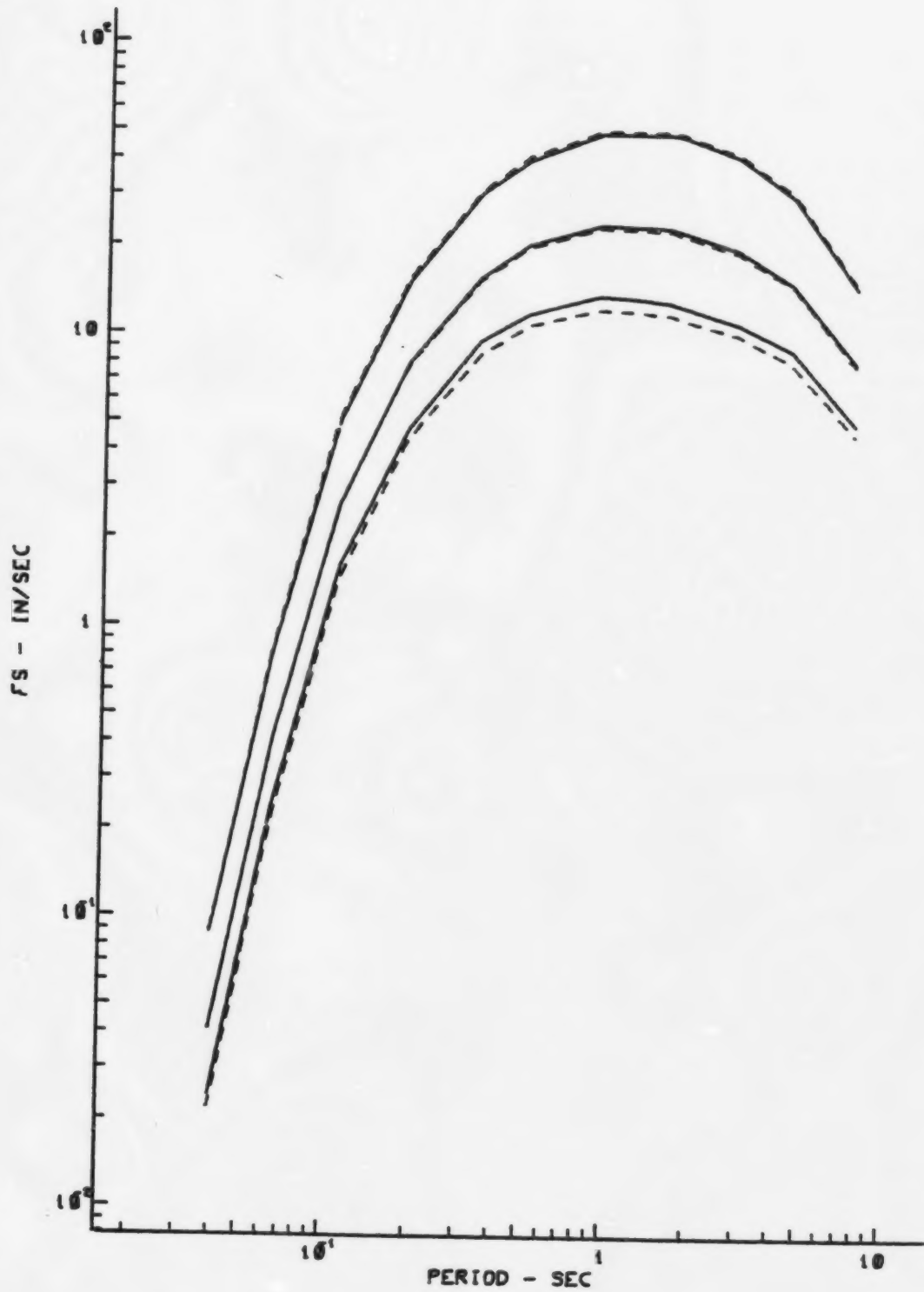


Figure 6.4.6

Finally, a variation of Model D is considered. Recall that for this it is assumed that since its eastern region (region II) did not have any large historic earthquakes, that that portion of the plate margin might be considered as a more likely candidate for a large earthquake in the future. One possible example is to assume then that the expected number of earthquakes in this eastern region (region II) will be literal, while leaving the expected earthquake occurrence rates of the western region (region I) to be Poissonian. Figure 6.4.6 shows the uniform risk spectra for this model at the 10, 50 and 90% probability levels of exceedance. The solid lines represent the risk spectra with the seismicity in the eastern region assumed to be literal while the dashed lines represent the assumption that all the seismicity in all the regions (I and II) is Poissonian.

ACKNOWLEDGEMENTS

This work was supported in part by a contract from the U.S. Nuclear Regulatory Commission through SEEC and by grants from the National Science Foundation.

REFERENCES

1. Algermissen, S. T. and Perkins, D. M. (1976). "A Probabilistic Estimate of Maximum Acceleration in Rock in the Contiguous United States," Department of the Interior Geological Survey, Open File Report 76-416.
2. Anderson, J. G. and Trifunac, M. D. (1977). "On Uniform Risk Functionals which Describe Strong Earthquake Ground Motion: Definition, Numerical Estimation and an Application to the Fourier Amplitude of Acceleration," Report No. CE 77-02, Civil Engineering Department, University of Southern California, Los Angeles.
3. Anderson, J. G. (1978). "Program EQRISK: A Computer Program for Finding Uniform Risk Spectra of Strong Earthquake Ground Motion," Report No. CE 78-11, Civil Engineering Department, University of Southern California, Los Angeles.
4. Anderson, J. G. (1979a). "Estimating the Seismicity from Geological Structure for Seismic-Risk Studies," Bull. Seism. Soc. Am., 69, pp. 135-158.
5. Anderson, J. G. (1979b). "On the Attenuation of Modified Mercalli Intensity with Distance in the United States," Appendix D, in Methods for Prediction of Strong Earthquake Ground Motion, U.S. Nuclear Regulatory Commission, NUREG/CR-068
6. Brune, J. (1970). "Tectonic Stress and the Spectra of Seismic Shear Waves from Earthquakes," J. Geophys. Res., 75, pp. 4997-5009.
7. Cornell, C. A. (1968). "Engineering Seismic Risk Analysis," Bull. Seism. Soc. Amer., 58, pp. 1583-1606.
8. Dalal, J. G. (1973). "Probabilistic Seismic Exposure and Structural Risk Evaluation," Ph.D. Dissertation, Stanford University.
9. DeCapua, N. J. and Liu, S. C. (1974). "Statistical Analysis of Seismic Environment in New York State," Fifth Symposium on Earthquake Engineering, Roorkee, India, pp. 389-396.
10. Der-Kiureghian, A. and Ang, A. H-S (1975). "A Line Source Model for Seismic Risk Analysis," University of Illinois, Urbana.
11. Der-Kiureghian, A. and Ang, A. H-S (1977). "Risk Consistent Earthquake Response Spectra," Proc. Sixth World Conference on Earthquake Engineering, New Delhi, India.
12. Der-Kiureghian, A. (1977). "Analysis of Uncertainties in Seismic Risk Evaluation," Proc. 2-nd ASCE EMD Specialty Conference, N. Carolina State Univ., N. Carolina, pp. 320-323.

13. Douglas, B. M. and Ryall, A. (1975). "Return Periods for Rock Acceleration in Western Nevada, Bull. Seism. Soc. Amer., 65, pp. 15-99-1611.
14. Gusev, A. A. (1983). "Descriptive Statistical Model of Earthquake Source Radiation and its Application to an Estimation of Short-Period Strong Motion," Geophys. J. Royal Astr. Soc., 74, pp. 787-808.
15. Kelleher, J., Sykes, L. and Oliver, J. (1973). "Possible Criteria for Predicting Earthquake Locations and their Application to Major Plate Boundaries of the Pacific and the Caribbean, J. Geophys. Res., 78, pp. 2547-2585.
16. Lee, V. W. and Trifunac, M. D. (1984). "Current Developments in Data Processing of Strong Motion Accelerograms," Report No. 84-01, Civil Engineering Department, University of Southern California, Los Angeles.
17. Lee, V. W. and Trifunac, M. D. (1985). "Attenuation of Modified Mercalli Intensity for Small Epicentral Distance in California, Report No. 85-01, Civil Engineering Department, University of Southern California, Los Angeles.
18. Liu, S. C. and Fagel, L. W. (1972). "Earthquake Environment for Physical Design: A Statistical Analysis," The Bell System Technical Journal, 51, pp. 1957-1982.
19. McGuire, R. K. (1974). "Seismic Structural Response Risk Analysis, Incorporating Peak Response Regressions on Earthquake Magnitude and Distance," Department of Civil Engineering, Massachusetts Institute of Technology, Cambridge MA.
20. Milne, W. G. and Davenport, A. G. (1969). "Distribution of Earthquake Risk in Canada," Bull. Seism. Soc. Amer., 59, pp. 729-754.
21. Oetken G., Parks, T. W. and Schuster, H. W. (1975). "New Results in the Design of Digital Interpolators," IEEE Trans. on Acoustics, Speech and Signal Processing, ASSP-23, No. 3, June.
22. Richter, C. F. (1958). "Elementary Seismology," Freeman Publishing Company, San Francisco.
23. Trifunac, M. D. (1976). "Preliminary Empirical Model for Scaling Fourier Amplitude Spectra of Strong Ground Motion in terms of Earthquake Magnitude, Source to Station Distance, and Recording Site Conditions," Bull. Seism. Soc. Amer., 66, pp. 1343-1373.
24. Trifunac, M. D. (1977). "Forecasting the Spectral Amplitudes of Strong Earthquake Ground Motion," Proc. Sixth World Conference on Earthquake Engineering, New Delhi, India.
25. Trifunac, M. D. and Anderson, J. G. (1977). "Preliminary Empirical Models for Scaling Absolute Acceleration Spectra," Civil Engineering Department, Report No. 77-03, University of Southern California, Los Angeles.

26. Trifunac, M. D. and Anderson, J. G. (1978a). "Preliminary Empirical Models for Scaling Pseudo Relative Velocity Spectra," Civil Engineering Department, Report No. 78-04, University of Southern California, Los Angeles.
27. Trifunac, M. D. and Anderson, J. G. (1978b). "Preliminary Empirical Models for Scaling Relative Velocity Spectra," Civil Engineering Department, Report No. 78-05, University of Southern California, Los Angeles.
28. Trifunac, M. D. and Lee, V. W. (1978). "Dependence of the Fourier Amplitude Spectra of Strong Motion Acceleration on the Depth of Sedimentary Deposits," Civil Engineering Department, Report No. 78-04, University of Southern California, Los Angeles.
29. Trifunac, M. D. (1979). "Preliminary Empirical Model for Scaling Fourier Amplitude Spectra of Strong Motion Acceleration in Terms of Modified Mercalli Intensity and Geologic Site Conditions," Int. J. Earthquake Engr. and Struc. Dyn., 1, pp. 63-79.
30. Trifunac, M. D. and Lee, V. W. (1979). "Dependence of Pseudo Relative Velocity Spectra of Strong Motion Acceleration on the Depth of Sedimentary Deposits," Civil Engineering Department, Report No. 79-02, University of Southern California, Los Angeles.
31. Trifunac, M. D. and Lee, V. W. (1985a). "Frequency Dependent Attenuation of Strong Earthquake Ground Motion," Civil Engineering Department, Report No. 85-02, University of Southern California, Los Angeles.
32. Trifunac, M. D. and Lee, V. W. (1985b). "Preliminary Empirical Model for Scaling Fourier Amplitude Spectra of Strong Ground Acceleration in Terms of Earthquake Magnitude, Source to Station Distance, Site Intensity and Recording Site Conditions: Second Paper," Civil Engineering Department, Report No. 85-03, University of Southern California, Los Angeles.
33. Trifunac, M. D. and Lee, V. W. (1985c). "Preliminary Empirical Model for Scaling Pseudo Relative Velocity Spectra of Strong Earthquake Acceleration in Terms of Earthquake Magnitude, Source to Station Distance, Site Intensity and Recording Site Conditions," Civil Engineering Department, Report No. 85-04, University of Southern California, Los Angeles.

NRC FORM 326 (2-84) NUREG 1100 2201, 3302		U.S. NUCLEAR REGULATORY COMMISSION		1 REPORT NUMBER (Assigned by NRC and Vol No., if any) NUREG/CR-4903, Vol. 2	
2 BIBLIOGRAPHIC DATA SHEET					
3 TITLE AND SUBTITLE Selection of Earthquake Resistant Design Criteria for Nuclear Power Plants - Methodology and Technical Cases: Methods for Introduction of Geological Data Into Characterization of Active Faults and Seismicity and Upgrading of the Uniform Risk Spectrum Technique					
4 DATE REPORT COMPLETED MONTH: September YEAR: 1986					
5 DATE REPORT ISSUED MONTH: May YEAR: 1987					
6 AUTHOR(S) J. G. Anderson, V. W. Lee and M. D. Trifunac					
7 PERFORMING ORGANIZATION NAME AND MAILING ADDRESS (Include Zip Code) Structural and Earthquake Engineering Consultants, Inc. Arcadia, California 91006					
8 PROJECT/TASK/WORK UNIT NUMBER					
9 PIN OR GRANT NUMBER FIN B8298 NRC-04-83-022					
10 SPONSORING ORGANIZATION NAME AND MAILING ADDRESS (Include Zip Code) Division of Engineering Office of Nuclear Regulatory Research U.S. Nuclear Regulatory Commission Washington, DC 20555					
11a TYPE OF REPORT Technical					
b PERIOD COVERED (Indicate dates) 1983 - 1986					
12 SUPPLEMENTARY NOTES					
13 ABSTRACT (200 words or less) This report reviews the physical and experimental bases for a quantitative relationship between earthquake occurrence rates and geological deformation rates. These relationships are well founded on mathematical statements of the elastic rebound theory, and well supported by observations. The concept of uniform risk spectra of Anderson and Trifunac (1977) is generalized to include (1) more refined description of earthquake source zones, (2) the uncertainties in estimating seismicity parameters a and b in $\log_{10} N = a - bM$, (3) the uncertainties in estimation of maximum earthquake size in each source zone, and (4) the most recent results on empirical scaling of strong motion amplitudes at a site. Examples of using the new NEQRISK program are presented and compared with the corresponding case studies of Anderson and Trifunac (1977). The organization of the computer program NEQRISK is also briefly described.					
14 DOCUMENT ANALYSIS - a KEYWORDS/DESCRIPTORS Earthquake Engineering Uniform Risk Spectra Strong Motion Seismology NEQRISK Computer Program					
b IDENTIFIERS/OPEN ENDED TERMS					
15 AVAILABILITY STATEMENT Unlimited					
16 SECURITY CLASSIFICATION (The page) Unclassified (The report) Unclassified					
17 NUMBER OF PAGES					
18 PRICE					